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Boris S. Sokolov

Academician, Russian Academy of Sciences
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The Sm–Nd Isotopic Systematics of Ophiolites in the Ozernaya Zone (Mongolia)

V. I. Kovalenko*, I. S. Pukhtel*, V. V. Yarmolyuk*, D. Z. Zhuravlev*, H. Stosch, and E. Jagoutz*****

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Abstract—The Sm–Nd isochron age was defined for rocks from various associations which make up part of the ophiolite assemblage of the Ozernaya zone of the early Caledonides. The age of gabbroids and tholeiite basic rocks of the silica-volcanic sequence in the Seriin-Nuru Range, which were formed under the conditions of an open marine basin, is 527 ± 43 Ma and $\varepsilon_{Nd}(T) = +6.7$. The Geriin–Nuru volcanics are associated with graywackes and were, most likely, generated in an island arc setting. Their age was found to be 522 ± 13 Ma, $\varepsilon_{Nd}(T) = +8.0$. The time of the regional metamorphic reworking of the ophiolites was determined on the basis of metamorphic minerals from rocks of the island-arc affinity of the Khalzan-Buregtei-Ula mountains and defined to be 487 ± 6 Ma with $\varepsilon_{Nd}(T) = +7.8$. The obtained data indicate the age similarity between MORB and island-arc ophiolites, and regionally metamorphosed rocks. They mark the final stage of the early Caledonian paleoocean with rapid change and lateral conjugation of various geodynamic settings.

Key words: *Ophiolites, metamorphism, isotopic age, Sm–Nd systematics.*

INTRODUCTION

Ophiolites have a significant role in the history of the fringing folded structures of the Siberian platform. They occur as basal formations of these structural zones and are commonly considered as the earliest manifestations of crust-forming processes within the region. In most cases, ophiolites are devoid of organic remains; therefore, their age position is defined on the basis of the age of the overlying rock units of the section or, more often, by the similarity of their geological position and composition with other ophiolites. Therefore, any direct evidence about the age of ophiolites is very important for correlating geological processes in the region. The article presents data on the Sm–Nd dating of the Ozernaya zone ophiolites, which is one of the largest in the south Caledonides fringing off the Siberian platform.

GEOLOGICAL CHARACTERISTICS

The ophiolitic Ozernaya zone is a part of the Great Lakes depression in western and southwestern Mongolia (Fig. 1). Abundant ophiolites occur here at the base of the stratified sequences (*Tektonika ...*, 1974). Ophiolite rocks commonly form tectonic blocks. Therefore, the general succession of magmatic events in the Ozernaya zone was reconstructed in the form of a composite section on the basis of the similarity of regularly repeated features (Dergunov, 1989) recorded in different blocks. The lower parts of the ophiolite section are presumably composed of rocks of the melanocratic assemblage (imbricated assemblages, represented by ultramafic rocks, gabbroids, and sheeted dikes). This assemblage is overlain by pillow basalts altered to greenstone. Basaltic lavas and diabase sills give way upsection to lava breccias, hyaloclastites, tuffosiliciliths, tuffites, and siliceous sediments. The carbonate–terrigenous–volcanic sequence, commonly enclosing basalts, andesites, dacites, and plagiorhyolites, occurs above the spilite–diabase complex. The sequence is rich in siliciclastic rocks, olistostromes, and reef-limestones. Recently, blocks of metamorphic rocks (gneisses and amphibolites) were discovered in the lower stream area of the left bank of the Kobdo River (the Ozernaya zone western border); according to data of the medium-scale geological survey, they are conventionally attributed to the Lower–Middle Ordovician, i.e. to the time of the main folding phase in the Ozernaya zone.

The samples we studied are from three levels of the above-mentioned composite section. The spilite–diabase pillow lavas, intercalated with siliceous sediments, the dikes cutting them, and the layered gabbroic sill were studied in the Seriin-Nuru Range, northeast of the Khoitu-Dalai-Nur (*S* in Fig. 1) in the central part of the Ozernaya zone. The tectonically bounded block is exposed here on the southern slope of the range among

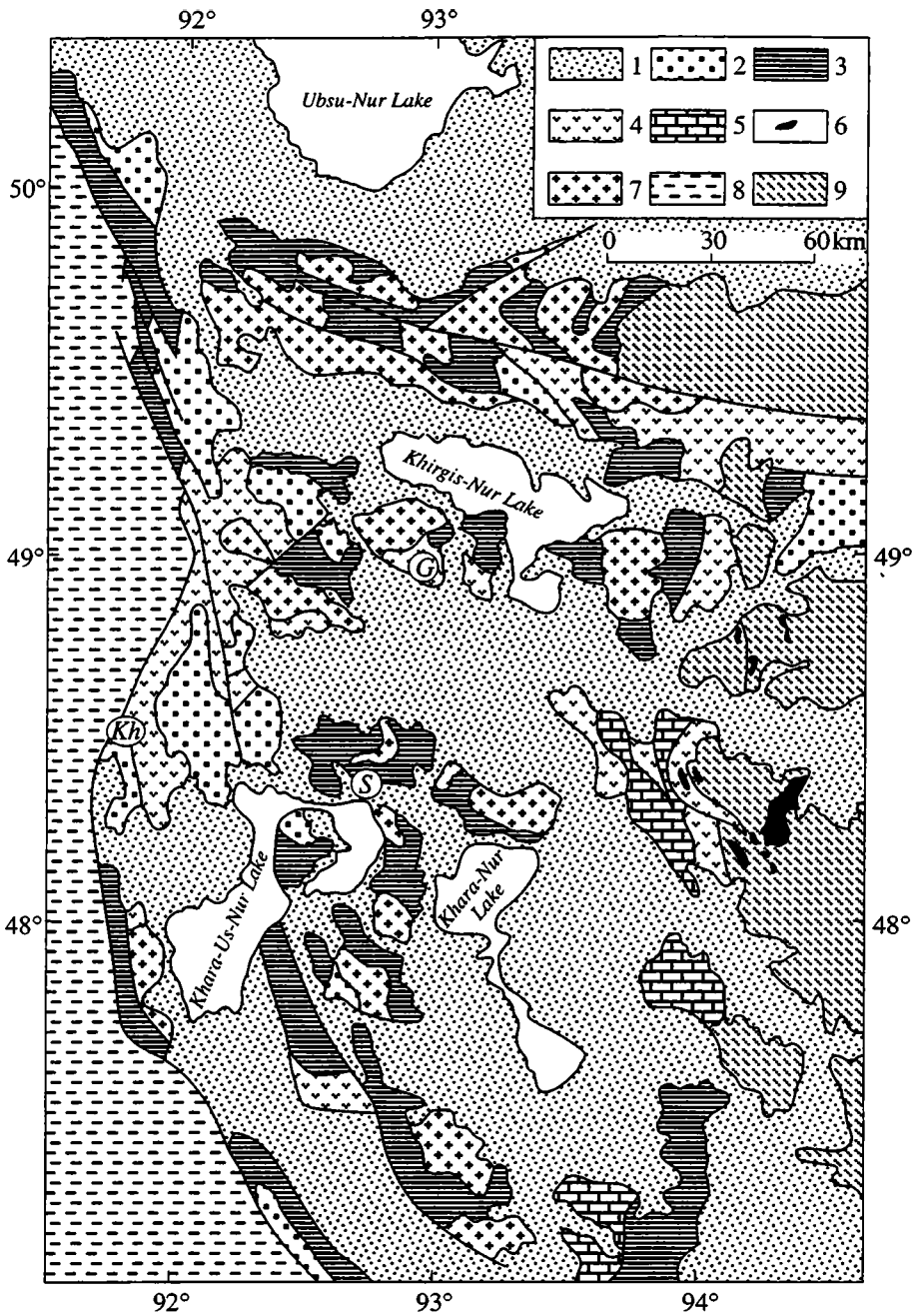

Fig. 1. Scheme of the geological structure of the Ozernaya zone of the early Caledonides in Mongolia complied using published map and scheme of tectonic zonation of Mongolia (*Karta* ..., 1989; *Geologicheskie* ..., 1995).

(1) Mesozoic–Cenozoic terrigenous complexes; (2) Middle Paleozoic complexes; (3–5) Vendian–Cambrian complexes: (3) ophiolite, (4) carbonate–terrigenous–volcanic, (5) terrigenous–carbonate; (6) allochthonous unltramafic rocks; (7) Early Paleozoic granitoids; (8) Caledonides of the Mongolia–Altai zone; (9) Precambrian crystalline complexes.

The sampling areas are indicate by letter symbols in circles: *S*—Seriin-Nuru, *G*—Geriin-Nuru, *Kh*—Khaldzan-Buregtei.

younger Devonian deposits; it is composed of rocks of the spilite–diabase facies and referred to the Vendian–Early Cambrian on the basis of geological evidence. Two fragments are identified in the block area we studied. They are separated by faults and exposed along the dry river bed near the Ielin-Khuduk Well (Fig. 2). The southern fragment is composed of greenstones after the pillow or massive basalts, interlayered with siliceous rocks, and cut by basite dikes and by the sill of stratified gabbroids. The thickness of lava flows is a few meters, and for sedimentary interlayers, about 0.5 m. The northern fragment of the block is represented by

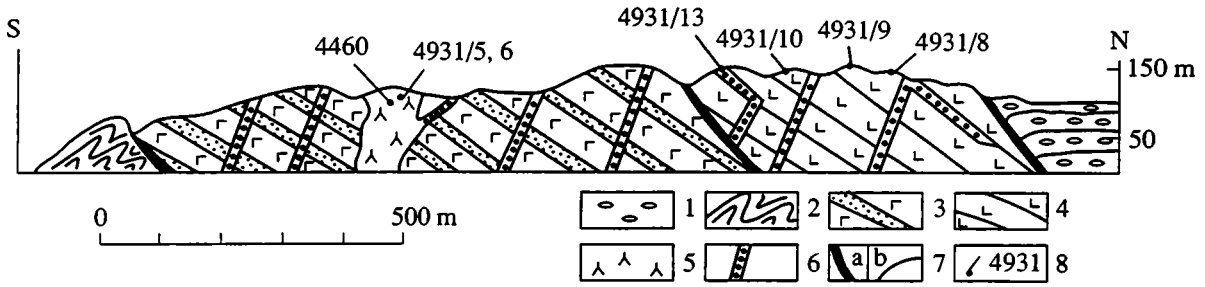

Fig. 2. Section through the ophiolite assemblage at the southern border of the Seriin-Nuru Range.
(1) red molasse (D); (2) sandy–siltstone deposits (D?); (3–7) ophiolite assemblage: (3) basalts, interbedded with siliceous deposits, (4) pillow and massive basalts, (5) layered gabbro, (6) basaltic dikes, (7) faults (a), geological boundaries (b); (8) sampling sites and their numbers.

homogenous pillow lavas enclosing rare thin interlayers of hyaloclastites and siliceous sedimentary rocks. The lavas are cut by basaltic dikes.

Samples for isotopic and geochemical analyses were collected from pillow lavas of the massive part of the section (SN-4931/8–10, Table 1), basaltic dikes (SN-4931/13), and a stratified body of gabbroids (SN-4460, SN-4931/5, 6, Table 1). All of the studied rocks, except gabbroids, correspond to spilites or spilitized basalts bearing albite (An_{17}). In addition to albite, phenocrysts are represented by clinopyroxene ($Wo_{36-43}En_{38-45}Fs_{18-21}$) and serpentinized olivine. The ground mass comprises glass replaced by actinolite and chlorite, plagioclase microlites, occasional round ore mineral segregations, carbonate, and epidote. Gabbroids are composed of plagioclase (from anorthite to andesine), clinopyroxene (Fe-index 0.25–0.31), and occasionally by titanomagnetite. High-Ti amphibole, which substitutes for the titanomagnetite outer rims, is recorded as an accessory and relatively late mineral. The rock texture is gabbroic. The sill layering is expressed as the alternation of more leucocratic (up to anorthosite) and more melanocratic rock layers with obscure boundaries, which are subparallel to the contacts of the body.

The sample GU-4898/5 is from the carbonate–terrigenous–volcanic sequence with basalts, andesites, and dacites. It was collected at the top of Mt. Gelben-Ula, the southern coast of the Khirgiz-Nur Lake (*G* in Fig. 1). Basalts, basaltic andesites and their tuffs with clinopyroxene and amphibole megacrysts are exposed here. In addition to these minerals, other phenocrysts are composed of plagioclase, clinopyroxene, hornblende, olivine, and magnetite. Olivine, clinopyroxene ($Wo_{45-47}En_{45-47}Fs_{7-9}$), and hornblende are mainly unaltered, whereas plagioclase phenocrysts and microlites are locally intensely albitized, and they have a basic composition only in relicts (An_{85}). The ground mass is chloritized and actinolitized, and comprises amygdales composed of epidote, actinolite, and rare quartz.

The schist sample KhBU-IX was collected at the left bank of the Kobdo River, in a dry river bed with the Khara-Yamar-Khuduk Well, west of the Khalzan-Buregtei-Ula (*Kh* in Fig. 1). It characterizes the metamorphic sequence convetionally assigned by the geological data to the Lower–Middle Ordovician. These schists represent the orthorocks which compose the intrusive body in gneisses, and are cut by dikes of amphibolites. The schist sample consists of quartz, plagioclase (An_{30-38}), biotite, muscovite, magnetite, ilmenite, garnet, and sphene. The chemical composition suggests the plagiogranite composition of its premetamorphic protolith (Table 1).

INVESTIGATION METHODS

Mineral fractions (from 40 to 100 mg) for isotopic investigations were hand separated at the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry of the Russian Academy of Sciences (IGEM). Minerals were analyzed at the Mineralogisch–Petrographisches Institut, Universität zu Köln, applying the Camebax electron microprobe with accelerating voltage (15 kV), a target current of 20 nA and a measurement time for each element 10 s. The bulk rock composition was studied at IGEM by the wet chemical method. The trace and rare earth elements were studied at the Mineralogisch–Petrographisches Institut, Universität zu Köln using the method of inductively coupled plasma spectrometry (ICPs). The chemical preparation of samples and analytical procedure are described in detail by Roelandts *et al.* (1988).

Sm–Nd isotopic investigations were carried out at the Max-Planck Institut in Mainz using a technique already published (Chauvel *et al.*, 1985), though somewhat modified. The ^{149}Sm + ^{150}Nd mixed spike was added to samples before decomposition. The sample powder (100–150 mg) and mineral fractions (50–70 mg) were decomposed in hermetic teflon beakers with the mixture of the HF + HNO_3 acids at a temperature of about 120°C during 3–5 days. After decomposition, the dry residue was dissolved in 6 N HCl and was again dried out. The procedure of chemical extraction of total REE, and the separation of Nd from Sm is identical to that described by White and Patchett (1984). The mass spectrometry analysis on Nd and Sm was performed

Table 1. The content of main and trace elements in ophiolitic rocks of the Ozernaya zone

Components	SN-4460	SN-4931/5	SN-4931/6	SN-4931/8	SN-4931/9	SN-4931/10	SN-4931/13	GU-4898/5	KhBU-IX
SiO_2	47.2	43.5	47.5	55.6	49.5	57.2	50.3	54.6	69.3
TiO_2	2.08	2.23	2.14	0.92	1.30	1.07	2.28	1.37	0.92
Al_2O_3	17.3	16.5	18.2	15.4	17.4	14.5	14.4	17.1	13.9
Fe_2O_3	5.20	5.75	2.32	4.63	4.60	7.09	5.93	2.58	0.00
FeO	5.38	5.98	6.70	4.20	7.17	4.37	5.95	5.57	2.65
MnO	0.18	0.22	0.16	0.18	0.19	0.19	0.18	0.13	0.15
MgO	4.10	4.59	4.29	4.81	4.58	3.96	4.46	4.41	1.92
CaO	9.64	10.8	8.87	4.23	3.12	2.23	6.24	7.09	3.12
Na_2O	4.04	2.31	4.14	7.28	6.12	5.54	4.70	4.85	4.19
K_2O	0.65	0.73	1.10	0.61	1.23	0.57	2.20	0.85	2.10
P_2O_5	0.16	0.45	0.39	0.18	0.21	0.15	0.39	0.47	0.15
H_2O	0.39	1.14	0.11	0.00	0.21	0.10	0.36	0.21	0.00
H_2O^+	3.43	5.08	3.32	1.72	4.00	2.60	2.41	0.47	1.21
F	0.09	0.09	0.07	0.16	0.07	0.05	0.09	0.12	0.11
CO_2	0.15	0.38	0.20	0.11	0.55	0.87	0.17	0.11	0.15
Total	99.94	99.72	99.46	99.87	100.25	100.34	100.16	99.92	99.90
Cr		22	16	17	12	2	22	88	3
V		335	300	321	326	379	329	349	27
Co		46	43	32	36	41	51	30	5.8
Ni		22	16	16	12	6	13	42	4
Cu		39	48	66	32	83	43	85	9
Zn		95	78	68	85	123	128	71	91
Zr	127	114	132	74	87	77	189	92	189
Nb	7	6	6	2	3	2	9	6	6
Y	27.0	27.1	29.5	23.2	26.7	26.8	40.8	24.9	28.9
Sc		27	25	33	33	38	32	31	12
Sr	276	332	487	175	378	137	452	440	408
Rb	6	5	11	6	8	5	17	12	18
La		9.94	11.3	4.78	4.57	4.65	19.0	11.7	17.0
Ce		25.4	28.4	12.8	14.0	13.6	44.8	26.3	42.4
Nd		17.7	19.3	9.82	11.0	10.4	27.7	16.1	27.7
Sm		4.77	5.07	3.02	3.37	3.28	6.89	3.90	6.56
Eu		1.60	1.72	1.01	1.17	1.11	2.26	1.34	2.31
Gd		5.14	5.55	3.64	4.06	3.93	7.19	4.40	6.31
Dy		4.84	5.26	3.78	4.41	4.30	7.05	4.27	5.37
Er		2.63	2.85	2.36	2.71	2.71	3.99	2.44	2.97
Yb		2.33	2.62	2.31	2.64	2.61	3.78	2.27	2.94
Lu		0.32	0.36	0.33	0.37	0.36	0.53	0.31	0.42

Note: The content of main components in wt %, trace elements in ppm.

using the Finnigan-MAT 261 under the statistic regime. The correction for Sm–Nd fractionation effects during the measurements was done by normalizing isotopic ratios relative to $^{147}Sm/^{152}Sm$ and $^{146}Nd/^{144}Nd$ = 0.7219. The measurements of the La Jolla Nd-standard isotopic composition during the experimental runs yielded the average result of $^{143}Nd/^{144}Nd$ = 0.511809 ± 20 (one σ; N = 17). The measurement results of the Nd isotopic composition in the samples were corrected with reference to the La Jolla value of $^{143}Nd/^{144}Nd$ = 0.511860 (Jacobson and Wasserburg, 1980, 1984). The calculation of isochron parameters was done by the York method (York, 1966). The error in the $\varepsilon_{Nd}(T)$ initial value was calculated using the Flecher and Rosman method (Fletcher and Rosman, 1982). While calculating the isochron parameters, the following errors were allowed for the isochron diagram coordinates for $^{147}Sm/^{144}Nd$—0.2%, and for $^{143}Nd/^{144}Nd$—the maximum divergence value between the reproducibility error in a series of La Jolla analyses (±0.000020)

and the error for individual analyses. Errors in the isochron parameters correspond to a 2σ or 95% confidence interval.

RESULTS AND DISCUSSION

According to the chemical composition (Table 1), the investigated samples of basic rocks are referred to gabbro (massive gabbroids), basalts, and basaltic andesites. Abundant albite in the volcanic rocks, together with their high Na_2O and low CaO contents, suggest that they are spilites or spilitized basalts.

It is not so easy to infer the initial composition of magmatic ophiolite rocks, in spite of numerous publications on methods concerning this problem. In our case, different methods do not bring identical results. The most reliable conclusions are as follows:

(1) Basalts and gabbroids of the Seriin-Nuru Range are associated with siliceous sediments, typical of open marine basins, and are characterized by the accumulation of iron and titanium in more differentiated rock varieties; the latter is typical of the tholeiitic magmatic series. The Geriin-Nuru rocks are weakly differentiated, so the iron distribution does not provide an opportunity to infer the particular petrochemical group of the rocks. The fact that among them there are varieties with amphibole phenocrysts and low-Fe dacites suggest that the rocks belong to the calc-alkaline island-arc series. The association of these rocks with graywacke sandstones and siltstones implies the proximity of the original setting to an island-arc.

(2) The REE distribution patterns for rocks of the Seriin-Nuru Range are of the intermediate type between MORB and island-arc tholeiite series. The Rb, Th, Hf, Zr, Ti, Y, Nb, and REE contents insignificantly (by a factor less than 2) deviate from the average contentrations of these elements in MORB (Sharas'kin, 1992). The island-arc volcanic rocks differ from MORB by the reduced content of Nb and Ta, and to a lesser extent, of Zr and Hf. Basic volcanics of the Seriin-Nuru Range show only Ta-depletion, though less prominent than in island-arc series.

(3) The REE distribution in samples SN-4931/13 and GU-4898/5 greatly differs from that of MORB, though it is similar to that of island-arc calc-alkaline volcanics.

(4) With their values of La/Nb and La/Th ratios, the studied rocks fall within the MORB field (Sharas'kin, 1992), or at the boundary between MORB and island arc rocks.

The obtained data on the REE distribution and geological association of the rocks under study suggests that the Seriion-Nuru basites are similar to MORB and marginal basin tholeiites with certain components of island-arc origin, which is quite typical of magmatic products in spreading zones of marginal basins. Sample GU-4898/5 seems to be close to island-arc calc-alkaline basalts.

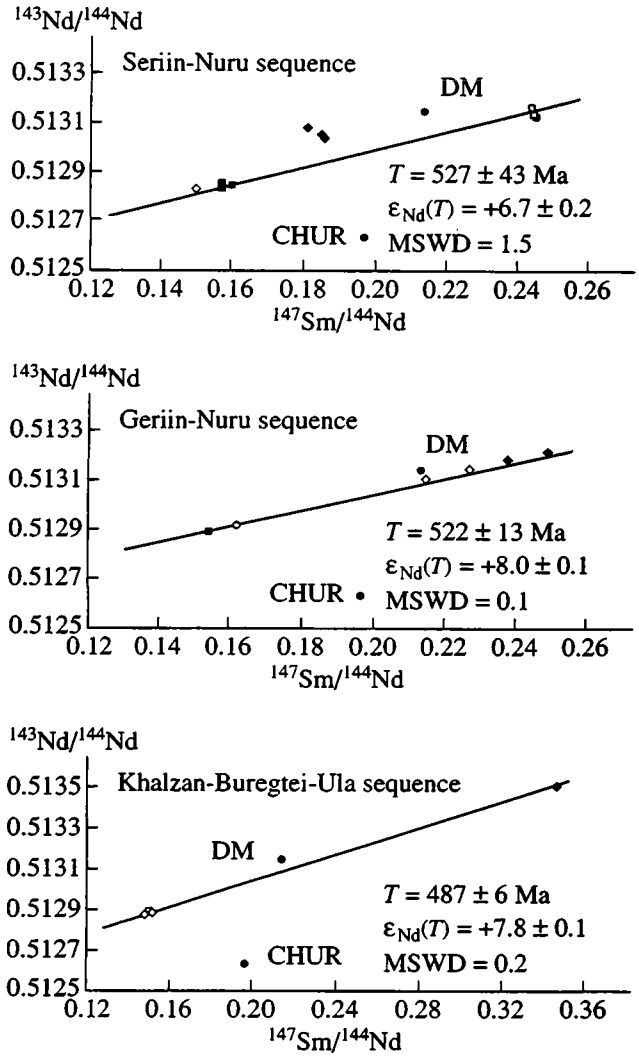

Fig. 3. Sm–Nd evolution diagrams for the Ozernaya zone ophiolites.

The Seriin-Nuru sequence: filled and open squares— whole rock samples and clinopyroxene fractions from gabbro sill; filled and open rhombs—basalts 4931/8–10 and 4931/13 respectively;

The Geriin-Nuru sequence: square—whole rock sample, open circle—olivine, filled and open rhombs—clinopyroxenes and amphiboles respectively;

The Khalzan-Buregtei-Ula sequence: square—whole rock sample, filled and open rhombs—garnet and plagioclase and biotite respectively.

The results of the Sm–Nd isotopic systematics for whole rock samples and mineral fractions from all of the above-discussed sequences (Seriin-Nuru, Geriin-Nuru, and Khalzan-Buregtei-Ula) are shown in Table 2 and in the Sm–Nd evolutionary diagrams (Fig. 3).

The Seriin-Nuru sequence. The data points for whole rock samples gabbro and clinopyroxene fractions fit the isochron age of 527 ± 43 Ma, indicating the emplacement time of sill with a rather high positive value of $\varepsilon_{Nd}(T) = +6.7 \pm 0.2$ (MSWD = 1.5). The isotopic parameters of the basaltic dike sample 4931/13

Table 2. Sm–Nd isotopic data

Sample no.	Rock	Sm, ppm	Nd, ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\varepsilon_{Nd}(T)$*
Seriin-Nuru sequence						
SN-4460 WR	Gabbro	5.173	19.82	0.15779	0.512841 ∓ 8	+6.6
SN-4460 Cpx		4.755	11.74	0.24488	0.513137 ∓ 6	+6.5
SN-4931/5 WR	Gabbro	4.576	17.31	0.15987	0.512851 ∓ 9	+6.6
SN-4931/5 Cpx		4.497	11.15	0.24377	0.513166 ∓ 9	+7.1
SN-4931/6 WR	Gabbro	5.002	19.17	0.15777	0.512862 ∓ 7	+7.0
SN-4931/6 Cpx		4.366	10.80	0.24439	0.513141 ∓ 8	+6.6
SN-4931/8 WR	Basalt	2.955	9.870	0.18096	0.513083 ∓ 25	+9.8
SN-4931/9 WR	Basalt	3.326	10.89	0.18472	0.513055 ∓ 13	+9.0
SN-4931/10 WR	Basalt	3.150	10.26	0.18562	0.513043 ∓ 9	+8.7
SN-4931/13 WR	Basalt	6.643	26.71	0.15035	0.512835 ∓ 7	+7.0
Geriin-Nuru sequence						
GU-4898/5 WR	Basalt	4.710	18.44	0.15438	0.512902 ∓ 8	+8.0
GU-4898/5 Ol		1.815	6.759	0.16234	0.512928 ∓ 10	+7.9
GU-4898/5 Cpx1**		4560	11.60	0.23771	0.513190 ∓ 17	+8.0
GU-4898/5 Cpx2***		4.465	10.85	0.24893	0.513220 ∓ 13	+7.9
GU-4898/5 Amph1**		1.701	4.788	0.21478	0.513109 ∓ 29	+8.0
GU-4898/5 Amph2***		1.347	3.593	0.22665	0.513156 ∓ 32	+8.1
Khalzan-Buregtei-Ula sequence						
KhBU-IX WR	Gneiss	7.002	28.12	0.15056	0.512891 ∓ 15	+7.8
KhBU-IX Gar		7.055	12.35	0.34538	0.513510 ∓ 9	+7.8
KhBU-IX Plag		2.925	11.68	0.15141	0.512892 ∓ 18	+7.8
KhBU-IX Bi		1.140	4.627	0.14902	0.512880 ∓ 9	+7.7

Note: * The ε_{Nd} values are recalculated for the isochron age T of each sequence. Mineral fractions were treated in 2 N (**) and 4 N HCl (***) respectively. WR, Ol, Cpx, Amph, Gar, Plag, Bi indicate the whole rock and olivine, clinopyroxene, amphibole, garnet, plagioclase, and biotite fractions, respectively.

[$\varepsilon_{Nd}(T)$ = +7.0] approaches those in the investigated massive gabbroids. Pillow lava samples (mostly spilites) from the massive part of the spilite–diabase sequence have much higher values of the Sm/Nd ratios and a more radiogenic Nd isotope composition [$\varepsilon_{Nd}(T)$ value varies from +9 to +10].

The linear approximation of analytical data for the lava sample 4898/5 from the Geriin-Nuru sequence and its clinopyroxene and amphibole fractions yielded the age of 522 ± 13 Ma, i.e. identical to the age of gabbro sill. The Nd isotope composition shows that this rock is intermediate between the gabbro sill and volcanic rocks of the Geriin-Nuru locality with more radiogenic Nd [$\varepsilon_{Nd}(T) = +8.0 \pm 0.1$].

The whole rock sample KhBU-IX from the Khalzan-Buregtei-Ula and metamorphic mineral fractions identify in the Sm–Nd evolutionary diagram the line with the slope, corresponding to the age of 487 ± 6 Ma [$\varepsilon_{Nd}(T) = +7.8 \pm 0.1$].

Geological data do not suggest any age difference between basalts of the spilite–diabase sequence (sample SN-4931/8–10) and gabbroids and basalts of the dike group. The calculated high $\varepsilon_{Nd}(T)$ values of +9 to +10 seem to agree with the geological and geochemical data, which indicate that these rocks were formed under conditions of oceanic spreading. Taking into consideration the lower value of $\varepsilon_{Nd}(T) = +7$ for gabbro, we have to admit that the island-arc component was involved in the process of magma generation and formation of the spilite–diabase sequence.

The obtained age values for ophiolitic magmatic rocks in the Ozernaya zone are very similar (within the measurements error) to the rocks of the Bayan–Khongor zone (569 ± 21 Ma; Kepezhinskas *et al.*, 1991), but substantially younger than the Bayanur complex of the Daribi Range (695 ± 25 Ma) in the Ozernaya zone (Khain *et al.*, 1995). Therefore, ophiolites in the Ozernaya zone may be different in age, suggesting a long time interval of its evolution.

The Sm–Nd age for the orthogneiss sample KhBU-IX from the Khalzan-Buregtei-Ula sequence corresponds to the time of regional metamorphism at the stage of general collision of the early Caledonides in the southern framing zones of the Siberian platform (Sklyarov, 1994).

Our results suggest that MORB, island-arc, and deeply metamorphosed ophiolite sequences are similar in age. This agrees with an assumption of A.N. Didenko *et al.* (1994) on the concurrent existence and, most likely, spatial conjugation of several spreading zones associ-

ated with large enough marginal seas, island arc systems, and zones of intense regional metamorphism within the Ozernaya zone.

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Triassic Stratigraphy of the Northern Caucasus Based on Ammonoids

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Abstract—Field investigations and bed-by-bed sampling of ammonoids provide the basis for a more detailed and better constrained biostratigraphic scheme of Triassic deposits in the northern Caucasus. For the first time, these deposits are subdivided into ten ammonoid beds comprising the Triassic interval from the middle Scythian to the upper Rhaetian. The beds are correlated with the standard ammonoid zones. The Anisian deposits, which are especially rich in ammonoids, are correlated with synchronous sedimentary sections in Turkey, Iran, China, and the USA. New data elucidate that in the studied, region there is a considerable stratigraphic hiatus spanning the upper Anisian and the lower Ladinian.

Key words: *Triassic, northern Caucasus, biostratigraphy, ammonoids, correlation.*

Triassic deposits are exposed in upper courses of the Malaya Laba and Belaya rivers, the southernmost part of the Krasnodar region. Along the strike, the maximum distance between their outcrops is less than 70 km. The main distribution area of the deposits is bounded by the Peredovoi Range. In a transitional zone between the Inzhenernyi Range and Mt. Pshekish spurs (near the Guzeripl'), Triassic formations are less widespread.

The Triassic deposits form a large anticline, the axis of which plunges to the northwest. Near the Novoprokhladnoe (south of the Kamennomostskii settlement), they are covered with Jurassic deposits. In general, the Triassic deposits transgressively overlap older rocks, such as Permian sediments, Paleozoic schists, and Precambrian gneisses. In turn, they are discordantly overlain by Jurassic formations.

In the northern Caucasus, Triassic deposits are represented by marine terrigenous-carbonate sediments. This sedimentary sequence shows breaks of various stratigraphic rank. The breaks are most significant in the eastern Peredovoi Range (Mt. Yatyrgvarta), where Norian conglomerates and sandstones overlie the lower Anisian limestones, and in the transitional zone (along the Belaya River), where the same Norian rocks overlap the Lower Permian deposits. The most complete Triassic sections occur in the central part of the Peredovoi Range, namely, on the Mt. Bol'shoi Tkhach, and along the Tkhach, Acheshbok, and Bugunzha rivers (Fig. 1), where they have a thickness of more than 1000 m.

The distribution of paleontological remains throughout these sections is irregular. Some horizons of the sections contain abundant and diverse fauna that enables them to be correlated with chronozones. Other members yield poor fossil assemblages. In these cases, the host deposits are correlative only to stratigraphic stages, or substages.

HISTORICAL REVIEW

V.I. Vorob'ev, the custodian of the Mineral Department at the Museum of Geology, Russian Academy of Sciences, was first to discover the Triassic deposits in the northern Caucasus. During the summer of 1906, he collected brachiopods and bivalves from outcrops of the Malaya Laba River. Studying this fauna later, Chernyshov (1907a, 1907b) and Borisyak (1909) determined the Late Triassic age of the host limestones. This was Vorob'ev's last expedition; he died after falling into a deep fracture of the Dzitaku glacier.

Later, the study on Triassic stratigraphy in the northern Caucasus was continued by Wittenburg and Robinson. In 1911, they visited the Malaya Laba and Belaya river basins and distinguished there sediments corresponding to all the Triassic stages (Wittenburg, 1912, 1913). During this expedition, Robinson collected several poor remains of ammonoids from platy limestones exposed southeast of the Mt. Bol'shoi Tkhach. After an expert examination by K. Diener, a professor at Vienna University, they were identified as Anisian forms. The ages of other beds have been constrained using bivalves and brachiopods. Carrying on further investigations, which were concentrated for almost 50 years in the Laba–Belaya interstream area, Robinson (1932, 1937, 1956) discovered ammonoids that characterize nearly all Triassic stages. Among them, 14 species were identified and described before the middle of this century

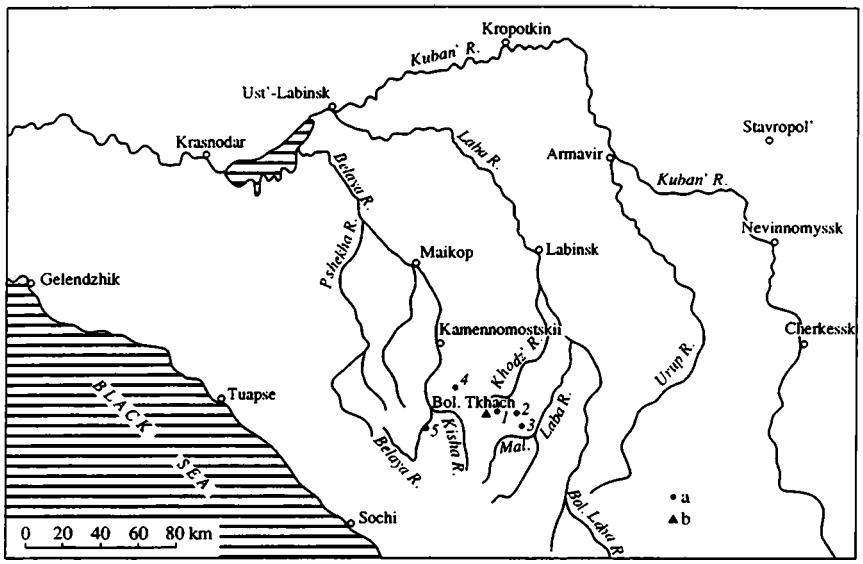

Fig. 1. Main localities of sections with Triassic ammonoids in the northern Caucasus. a – localities: *1*—Tkhach River; *2*—Bugunzha River; *3*—Mt. Yatyrgvarta; *4*—Sakhrai River; *5*—Settlement of Guzeripl'; b—Mt. Bol'shoi Tkhach.

(Robinson, 1936; Voinova *et al.*, 1947). Later, the Triassic ammonoids from the northern Caucasus were studied by Danilevich (1951), Shevyrev (1961, 1968), Popov (1962), and Kurgalimova (1977), who tripled the number of species in the list of specimens.

The first stratigraphic scheme for the Triassic of the northern Caucasus was suggested by Robinson (1932, 1936, 1937, 1956), who subdivided sections of this region into units corresponding to all stages of the Triassic System. In 1958, this scheme was approved by the First Interdepartmental Regional Stratigraphic Conference on the Mesozoic in the Caucasus.

Later, this region was thoroughly studied by several organizations, such as the Caucasus Expedition of Moscow State University (Dagys, 1963; Edigaryan, 1963; Slavin, 1964); the Krasnodar Division of the All-Union Petroleum-Gas Research Institute (Rostovtsev, 1960; Rostovtsev *et al.*, 1966); the Paleontological Institute, Russian Academy of Sciences (Shevyrev, 1968); and others. Results of these studies were published in two cumulative volumes: *Geology of the USSR* (Robinson, 1968) and *Stratigraphy of the USSR* (Dagys and Robinson, 1973). New data on Triassic stratigraphy in the northern Caucasus were compiled by a large group of scientists in *Ob"yasnitel'naya ...* (1973), edited by Kiparisova. A project involving a stratigraphic scheme was proposed by Rostovtsev (1973) in the latter publication; it was accepted by the Second Interdepartmental Regional Stratigraphic Conference on the Mesozoic in the Caucasus, May 17–20, 1977. Later, it was confirmed by the Interdepartmental Stratigraphic Committee (*Reshenie ...*, 1979). According to this scheme, the Triassic deposits in the northern Caucasus were subdivided into several groups and formations. After a revision, some paleontologists from Moscow suggested a modification of particular paragraphs of the scheme concerning the nomenclature, correlation, and age determinations of certain Triassic beds occurring in the northern Caucasus (Beznosov, 1975; Yaroshenko, 1978; Beznosov and Efimova, 1979).

My expeditions, performed over the past 25 years, also provided a substantial amount of new material. According to all available data, assemblages of Triassic ammonoids from the northern Caucasus are represented by 85 species (Shevyrev, 1995). This is a sound basis for improving the Triassic stratigraphy of the northern Caucasus, especially when it concerns Anisian subdivisions, for which a principally new scheme of correlation with standard zonation is suggested.

SUBDIVISIONS OF TRIASSIC DEPOSITS IN THE NORTHERN CAUCASUS

The Triassic deposits of the northern Caucasus are divisible into three units with different lithologic features: the lower group, comprising mostly carbonate rocks; the middle terrigenous group; and the upper carbonate group. These units are separated by stratigraphic hiatuses.

The lower unit has been distinguished as the Tkhach Group (Rostovtsev, 1973). It includes the Yatyrgvarta, Malyi Tkhach, and Acheshbok formations.

The Yatyrgvarta Formation is widespread in the Peredovoi Range. In its stratotype locality at Mt. Yatyrgvarta, the formation rests on Paleozoic gneisses and diorites (Fig. 2). It consists of three members. The lower member (30 m) comprises a basal conglomerate grading into coarse-grained sandstone. The middle member (25 m) is composed of thick beds of yellow-

gray limestones. The upper member (200 m) includes thin-platy limestones, the lower part of which yielded bivalves *Claraia clarai, C. stachei*, and *C. aurita*. The rich ammonoid assemblage from its middle part consists of *Juvenites sinuosus, Dieneroceras caucasicum, D. magnum, Preflorianites toulai, Meekoceras gracilitatis, M. caucasium, Arctoceras robinsoni, A. kiparisovae, Inyoites oweni, Subvishnuites welteri, Parussuria compressa, Lanceolites compactus, L. bicarinatus, Prosphingitoides austini, Owenites koeneni*, and *Subowenites slavini*. According to the dominant forms of the assemblage, I classify the host deposits as the *Owenites-Dieneroceras* Beds.

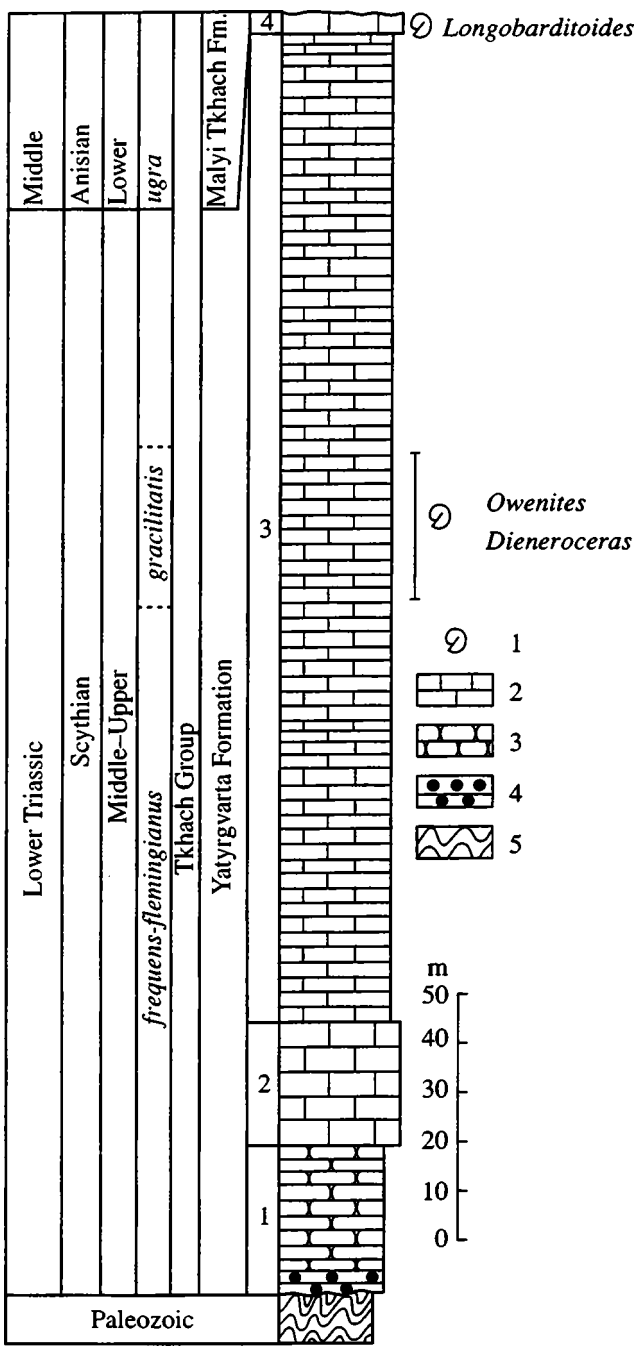

Fig. 2. Section of the Tkhach Group, Mt. Yatyrgvarta locality (the author's interpretation based on data of Robinson, 1932, and Rostovtsev, 1973).

1—ammonoids; 2—limestone; 3—sandstone; 4—conglomerate; 5—gneiss.

At Mt. Yatyrgvarta, the upper member is overlain by massive limestones of the Malyi Tkhach Formation, which is only 2 m thick here due to erosion. A more complete section of this formation is found in its stratotype locality, situated in the upper stream of the Tkhach River. Here, it is 50 m thick and consists of light gray massive limestones with pink-gray conglomerate interbeds (Fig. 3). Southward, in the Mertvaya Balka at Mt. Sunduki and in the Malyi Bambak and Skirda ranges, the formation is composed of lower massive, sometimes oolitic limestones associated with upper limy breccia and crinoidal limestones bearing brachiopods and rare ammonoids. In this region, it is 120–150 m thick. At the Mertvaya Balka locality, the Malyi Tkhach Formation yielded *Stenopopanoceras transiens, Groenlandites? glaber*, and *Megaphyllites* sp., while ammonoids collected from it at Mt. Shapka are represented by *Stenopopanoceras transiens, Longobarditoides caucasius, Laboceras gracile, Megaphyllites compressus, Leiophyllites pitamaha*, and *Longobarditoides caucasius*; the last form also occurs at Mt. Yatyrgvarta. I consider the deposits with this assemblage as the *Stenopopanoceras* Beds.

The overlying Acheshbok Formation is the main source of Triassic ammonoids collected in the northern Caucasus. In many localities, the contact between this and the Malyi Tkhach Formation is normal; however, in the upper courses of the Tkhach River, it rests on an erosion surface. In the stratotype locality of the Acheshbok River, the formation consists of two units: a lower subformation composed of limestones with shale interbeds and an upper shaly subformation enclosing limestone interbeds.

The most complete sections of the Acheshbok Formation occur in the central part of the region under discussion. Here, along the Tkhach River, its lower subformation is up to 110 m in thickness, and it consists of light gray, thin-platy, sometimes cloddy or lensing limestones with thin interbeds of dark gray limy shales and siltstones (Fig. 3). In the interval of 12–25 m above the base, it includes beds of gray sandstones, which are 1.5–3.0 m in thickness. The subformation yielded two ammonoid assemblages.

The first assemblage was collected from the member of cloddy limestones and shales occurring within the interval of 50–60 m above the subformation base. The assemblage includes *Alanites visendus, Parasageceras tkhachense, Sageceras* sp., *Norites labensis, Sturia sansovinii, Procladiscites elegans, Mesocladiscites caucasius, Caucasochordiceras simplex, Pseudohollandites densicostatus, Robinsonites caucasius, Paradanubites palmatus, P. orbiculatus, P.*

inornatus, Laboceras gracile, L. acutulum, Longobarditoides caucasius, Japonites labaensis, Caucasites evolutus, C. inflatus, Gymnites sp., *Megaphyllites prometheus, M. compressus, Leiophyllites pitamaha, L. visendus, L. ellipticus*, and *Monophyllites kieperti*. This assemblage, which characterizes the *Laboceras-Megaphyllites* Beds, was collected mainly along the left headstream of the Tkhach River. Some representatives, including additional forms, were also encountered in other localities of the central region situated along the Acheshbok River (*Aegeiceras byzovae* and *Megaphyllites prometheus*) and the Bugunzha River (*Gymnites* sp., *Megaphyllites prometheus, Leiophyllites visendus,* and *Monophyllites wengensis*).

The second assemblage of ammonoids was recovered from alternating limestones, shales, and siltstones, which compose the 14-m-thick member at the level 7 m below the top of the subformation. Hard limestones of the member form either 8-cm-thick beds or lenses (60 cm long) and concretions, which are extremely rich in spheroid shells of *Isculites sphaericus*, associated with rare *Sturia sansovinii, Megaphyllites compressus*, and *Leiophyllites visendus*. I identify this level as the *Isculites* Beds. They are exposed along the right tributary of the Tkhach River, upstream from the bridge. The *Isculites* forms are also known from outcrops of the Malyi Sakhrai and Bzhebs rivers.

The upper subformation is well-exposed along the Tkhach River, beginning from the bridge downstream, where it has a sharp contact with deposits of the Sakhrai Group. The subformation (50 m) consists of dark gray shales enclosing either thin-bedded or lumpy and cloddy limestones (Fig. 3). In its upper part, which bears ferri-calcareous concretions, the limestone interbeds are more rare than in the lower portion. Both the shales and the limestones of this subformation yielded

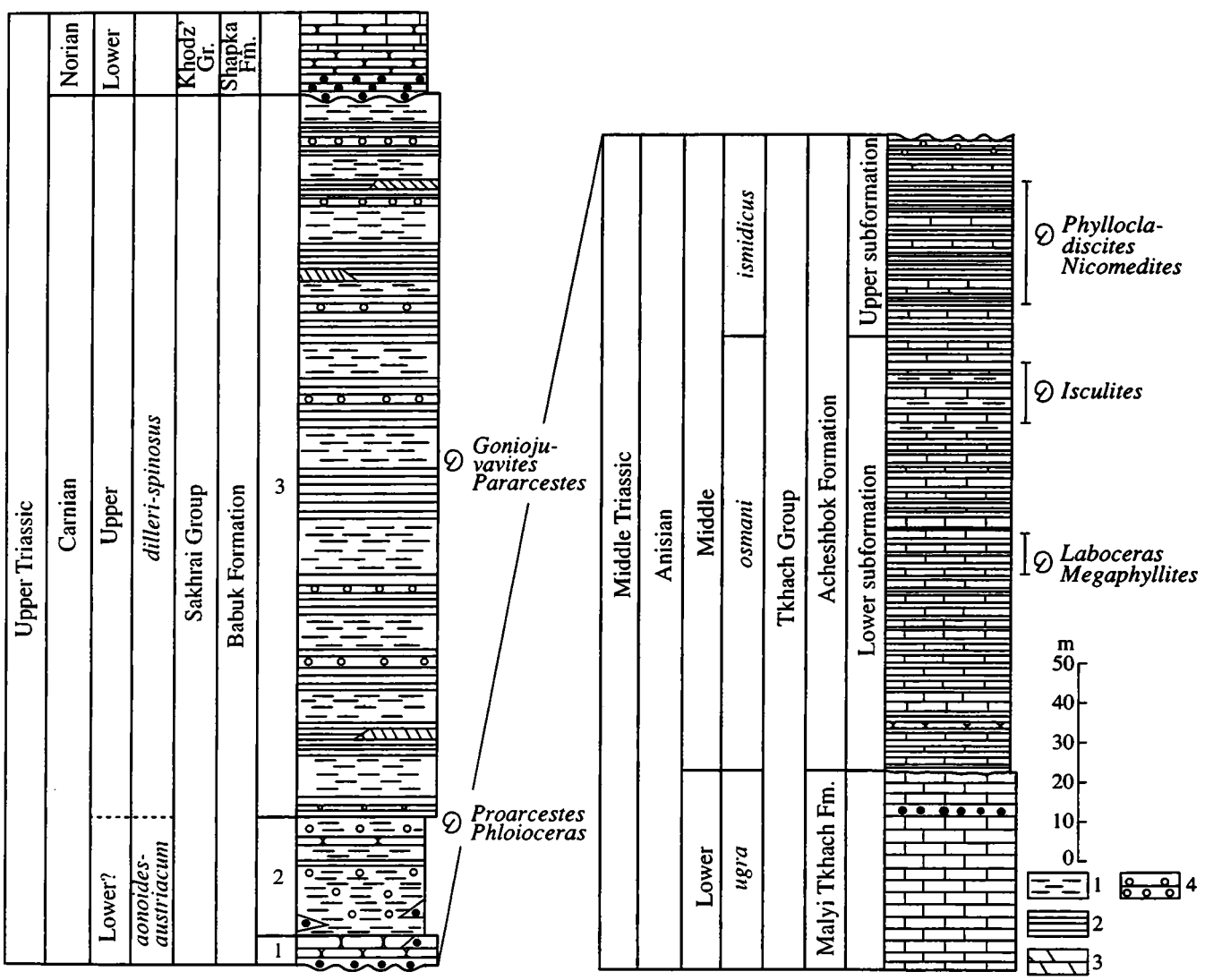

Fig. 3. Section of the Tkhach and Sakhrai groups, Tkhach River locality.
1—siltstone; 2—shale; 3—marl; 4—concretions; other symbols as in Fig. 2.

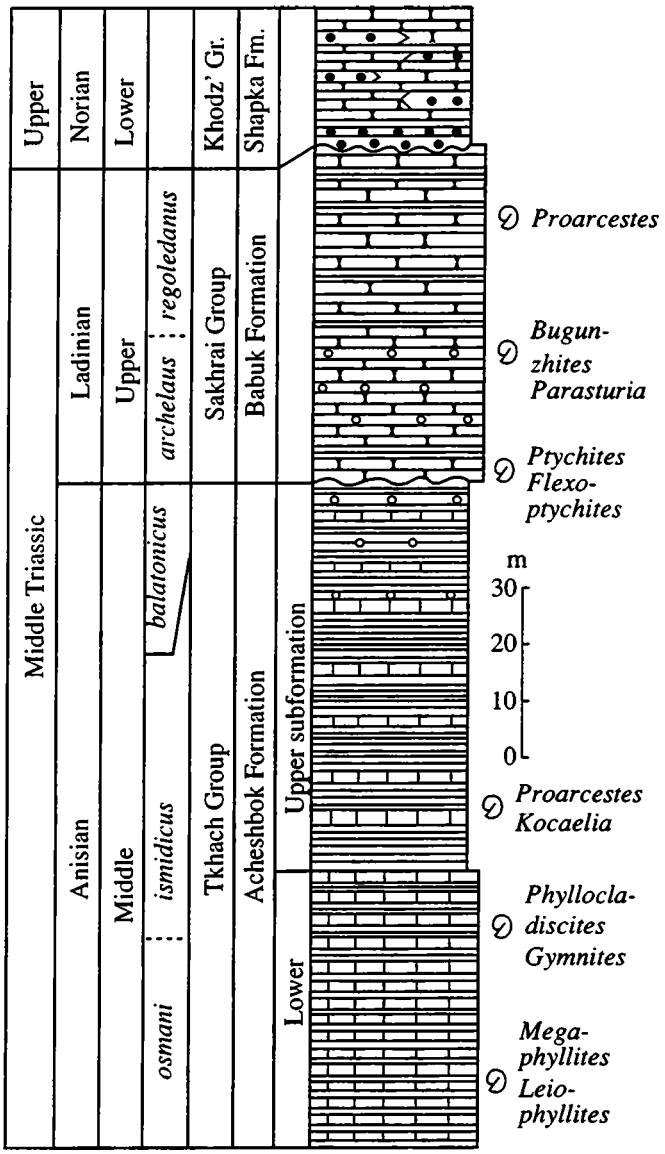

Fig. 4. Section of the Acheshbok Formation, Bugunzha headstream locality (the author's interpretation based on data of Beznosov, 1975; symbols as in Figs. 2 and 3).

the following ammonoids: *Vorobyevites unicus, Sturia sansovinii, Phyllosphaerites obscurus, Procladiscites elegans, Phyllocladiscites proponticus, Psilocladiscites* sp., *Acrochordiceras subrotundum, Epacrochordiceras inflatum, E. compressum, Nicomedites caucasius, N. tkhachensis, Kocaelia lenticularis, Longobarditoides magnus, Gymnites robinsoni, G. evolutus, Leiophyllites visendus, Monophyllites kieperti*, and *M. caucasius*. These forms are scattered throughout the entire interval of the subformation, and I consider this unit as the *Phyllocladiscites-Nicomedites* Beds. Some representatives of this rich assemblage have also been encountered in outcrops of the Acheshbok River (*Sturia sansovinii, Phyllocladiscites proponticus*, and *Gymnites robinsoni*), in the upper courses of the Bugunzha River (*Phyllocladiscites proponticus, Nicomedites caucasius, Kocaelia lenticularis, Gymnites robinsoni*, and *Proarcestes bramantei*), in the Mertvaya Balka (*Sturia sansovinii, Nicomedites caucasius, Kocaelia lenticularis*, and *Leiophyllites visendus*), along the Bzhebs River (*Nicomedites caucasius*), in the Svinyach'ya Balka, tributary of the Sakhrai River (*Proarcestes bramantei*), and Sakhrai River (*Epacrochordiceras compressum*).

In the Tkhach River localities, the upper part of the Acheshbok Formation is eroded. This stratigraphic interval, preserved in outcrops of the Bugunzha headstream, is represented by shales with rare limestone interbeds and ferri-calcareous concretions (Fig. 4). Beznosov collected from these rocks *Ptychites besnosovi, Flexoptychites bugunzhensis, Epacrochordiceras* sp., *Sturia* sp. and *Gymnites* sp. This ammonoid assemblage is typical of the *Ptychites-Flexoptychites* Beds. The forms *Kocaelia lenticularis* and *Proarcestes bramantei* were recovered from deposits below these beds, and their host rocks correspond to the *Phyllocladiscites-Nicomedites* Beds of the Tkhach River localities. In the Bugunzha upper courses, the total thickness of the upper Acheshbok Subformation is up to 80 m. The *Ptychites-Flexoptychites* Beds crown the Acheshbok Formation, and they represent the uppermost unit of the Tkhach Group.

The middle terrigenous unit of Triassic deposits in the northern Caucasus is known as the Sakhrai Group. It was originally distinguished as a formation of the same name comprising conglomerate, sandstone, and siltstone beds, which are exposed in the lower courses of the Sakhrai (Slavin and Robinson, 1962). Later, the name was used to identify all terrigenous rocks occurring between the two groups of calcareous deposits (Rostovtsev, 1973; *Reshenie* ..., 1979). Recent data show that the Sakhrai Group comprises units that differ in their ages (from the Anisian to the Carnian) and are separated by hiatuses (Beznosov, 1975; Yaroshenko, 1978; Beznosov and Efimova, 1979). The stratigraphic nomenclature for these units needs to be revised.

The first attempt at revision was made by Beznosov (1975), who suggested that part of the Sakhrai deposits be considered as the Babuk Formation. The outcrops of this formation are traceable from the Bugunzha-Acheshbok watershed to the Kuna River, and they also occur along the Tkhach and Malyi Sakhrai rivers and in the Svinyach'ya Balka. These deposits unconformably overlie various horizons of the Acheshbok Formations, and they are discordantly overlapped by the basal member of the upper calcareous group. In its upper part (40–50 m), the Babuk Formation consists of coarse clastic rocks that grade upward into shales and siltstones (150–200 m).

A representative section of this formation is exposed along the Tkhach River. The section begins with a member (7 m) of massive sandstone enclosing conglomerate lenses and resting on the eroded surface of the Acheshbok Formation (Fig. 3). The member is

overlain by dark gray siltstone (30 m) with rare sandstone interbeds, conglomerate lenses, and horizons of ferri-calcareous concretions. The top part of the siltstone yielded *Proarcestes planus* and nautiloid *Phloioceras* sp., and I identified this unit as the *Proarcestes-Phloioceras* Beds. The rest of the Babuk Formation (180 m) consists of gray micaceous siltstones alternating with dark gray shales and enclosing marl lenses and ferri-calcareous concretions. From the middle horizons of this unit exposed along the Tkhach River and on the eastern slope of Mt. Bol'shoi Tkhach, I collected *Clionitites* sp., *Goniojuvavites planus, Proarcestes planus*, and *Pararcestes acutus*. In addition, the same level yielded the following forms: *Projuvavites caucasius* and *Proarcestes* sp. at the Malyi Sakhrai locality; *Proarcestes planus* from the Sakhrai River outcrops; and *Goniojuvavites planus, Proarcestes wittenburgi, Pararcestes optimus, P. paropertus*, and *P. sublabiatus* from the Svinyach'ya Balka section. These ammonoids characterize the *Goniojuvavites-Pararcestes* Beds.

A peculiar section of the Sakhrai Group is exposed in the upper courses of the Bugunzha River (Fig. 4). Here, the group is 50–70 m thick, and it consists of gray fine-grained sandstones alternating with dark gray shales bearing ferri-calcareous concretions. Its lower part encloses *Bugunzhites dagysi, Parasturia acutata, Sturia semiarata*, and *Monophyllites wengensis*; the upper one yielded *Proarcestes* sp. Accordingly, I distinguished the lower part as the *Bugunzhites-Parasturia* Beds. In other regions of the Caucasus, these beds have not been encountered. In the described locality, they rest on the Acheshbok Formation without any evident unconformity, and Beznosov (1975) has suggested that they compose the uppermost part of the formation (the third subformation). However, this unit significantly differs from the Acheshbok Formation in both its lithologic and biostratigraphic characteristics, and this points to a latent hiatus at the base of the beds. Therefore, the beds belong to the Sakhrai Group, rather than the Acheshbok Formation. In the region of the Bugunzha headstream, the eroded surface of this unit is overlain by lensing conglomerates and sandstones with rare shale interlayers. These rocks compose the basal unit (20–30 m) of the upper calcareous group, i.e., the Shapka Formation (Beznosov and Efimova, 1979).

The upper sequence of calcareous rocks has been distinguished as the Khodz' Group (Rostovtsev, 1973; *Reshenie* ..., 1979). It transgressively overlaps all underlying deposits: the Babuk Formation (the Tkhach and Bugunzha rivers), the Malyi Tkhach Formation (Mt. Yatyrgvarta), and the Lower Permian rocks (the village of Guzeripl'). The Khodz' Group consists of two distinct units: the lower, dominantly terrigenous formation and the upper one, which is composed of carbonates. The lower unit, named the Shapka Formation (from 60–80 to 140 m in thickness), consists of conglomerate, sandstone, and sandy limestone beds. The upper unit, or Khodz' Formation (up to 300 m in thickness), is composed of pure limestones (Beznosov and Efimova, 1979). The erosion surface at the top of the Khodz' Formation underlies the Lower Jurassic (Lotharingian) conglomerates and sandstones.

The biostratigraphy of the Khodz' Group is well-studied in the section (Fig. 5) exposed on the western slope of Mt. Yatyrgvarta, where the beds above the massive limestones of the Malyi Tkhach Formation are represented by the following rocks:

1. Gravelstone and sandstone with basal conglomerate (5 m).

2. Platy sandstone (20 m).

3. Alternating sandy limestones and sandstones (15 m) with bivalves (*Monotis kaukasica*) and brachiopods (*Oxycolpella robinsoni* and *Rhaetina elliptica*).

4. Sandstone with gravelstone and small-pebble conglomerate interlayers (25 m).

5. Sandy limestone (12 m) with coral bioherms, bivalves (*Myophoria verbecki* and *Indopecten glabra*), and brachiopods (*Oxycolpella robinsoni, Rhaetina turcica, Rh. elliptica, Triadithyris gregariaformis*, and *Crurirhynchia kiparisovae*).

6. Red organoclastic limestone (40 m) with algal bioherms and abundant brachiopods (*Crurirhynchia kiparisovae, Moisseievia* cf. *moisseievi, Oxycolpella robinsoni, Pexidella strohmayri, Zeilleria moisseievi, Z. bukowskii, Rhaetina caucasica*, and others).

7. Red clayey limestone (10–15 m) with lenses of organoclastic limestones and abundant brachiopods (*Moisseievia skirdaensis, Euxinella iatirgvartaensis, Austrirhynchia cornigera, Zugmayerella koessenensis, Laballa suessi, Sinucosta bittneri, Wittenburgella minuta*, and *Rhaetina pyriformis*).

8. Dark gray sandy limestone (15 m) with sandstone and siltstone interlayers.

9. Alternating dark gray sandy and red clayey limestones (8–10 m) with brachiopods (*Euxinella robinsoni, E. lavantina, Oxycolpella oxycolpos, Majkopella worobievi, M. manzavini, Neoretzia superbescens, Laballa suessi, Zugmayerella koessenensis*, and *Rhaetina pyriformis*) and ammonoids (*Placites polydactylus, Megaphyllites insectus, Stenarcestes leiostracus*, and *Rhacophyllites debilis*).

10. Red and brown limestones (25–30 m) with brachiopods (*Robinsonella mastakensis, Oxycolpella oxycolpos, Majkopella worobievi, Laballa suessi, Zugmayerella koessenensis*, and *Neoretzia superbescens*).

Above, there are beds of Lower Jurassic conglomerates and sandstones. The Shapka Formation comprises members 1–4 of the section, while other units represent the Khodz' Formation. Member 9 is the only one in this section that yielded ammonoids, and I consider it to represent the *Placites-Rhacophyllites* Beds. Ammonoids of this stratigraphic level have also been encountered in other localities, such as the Baronovy Polyany (*Paracladiscites multilobatus, Megaphyllites insectus, Stenarcestes leiostracus*, and *Rhacophyllites debilis*), Mt. Bol'shoi Tkhach (*Placites polydactylus* and *Mega-*

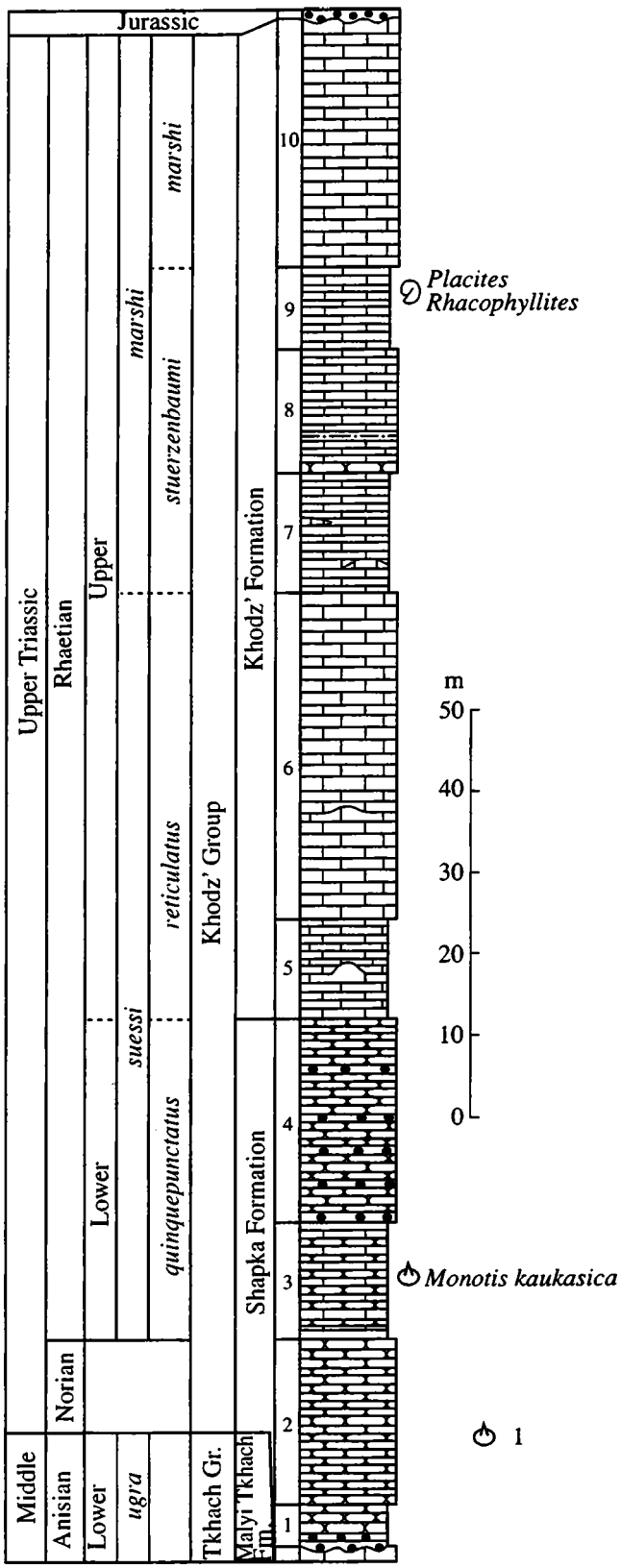

Fig. 5. Section of the Khodz' Group, Mt. Yatyrgvarta locality (the author's interpretation based on data of Dagys, 1974).
1—bivalves; other symbols as in Fig. 2.

phyllites insectus), the Kuna headstream area (*Paracladiscites multilobatus, Placites polydactylus, Megaphyllites insectus, Stenarcestes leiostracus*, and *Rhacophyllites debilis*), Bzhebs River (*Adygeites obscurus*), and the Malyi Sakhrai and Acheshbok rivers (*Placites polydactylus*).

A peculiar section of the Khodz' Group is exposed on the left bank of the Belaya River near the Guzeripl' (Fig. 6). The deposits, which rest here on an erosion surface of red-brown thick-bedded sandstones and conglomerates of the Lower Permian, are represented by the following members:

1. Light gray massive sandstone (90–100 m) with quartz-conglomerate interlayers.

2. The same rock (50 m) with two beds (8–12 m in thickness each) of dark gray massive limestones that are very rich in brachiopods: *Septaliphoria fassicostata, Rimirhynchopsis triadica, Oxycolpella oxycolpos, Rhaetina gregaria, Rh. pyriformis, Zeilleria bukowskii*, and others. The lower bed encloses an interlayer of very hard limestone (1 to 2 m) with ammonoids: *Cladiscites tornatus, Placites polydactylus, Pinacoceras metternichi*, and *Megaphyllites insectus*.

INTERREGIONAL CORRELATION

Triassic ammonoids from the northern Caucasus include many endemic species, and the interregional correlation of their host deposits is somewhat problematic. In this case, it seems reasonable to use their generic affinity, because the Triassic genera of ammonoids mostly had wide areal distributions, but narrow stratigraphic ranges.

As was shown, the Triassic sections of the northern Caucasus comprise 10 biostratigraphic horizons bearing different ammonoid assemblages. These are the *Owenites-Dieneroceras, Stenopopanoceras, Laboceras-Megaphyllites, Isculites, Phyllocladiscites-Nicomedites, Ptychites-Flexoptychites, Bugunzhites-Parasturia, Proarcestes-Phloioceras, Goniojuvavites-Pararcestes*, and *Placites-Rhacophyllites* beds.

The *Owenites-Dieneroceras* Beds of the Yatyrgvarta Formation represent the lowermost stratigraphic level of Triassic ammonoids from the Caucasus. Their rich and diverse assemblage includes the following 16 species of 12 genera: *Juvenites sinuosus* (Kiparisova), *Dieneroceras caucasicum* Popow, *D. magnum* Shevyrev, *Preflorianites toulai* (Smith), *Meekoceras gracilitatis* White, *M. caucasium* Shevyrev, *Arctoceras robinsoni* (Kiparisova), *A. kiparisovae* Shevyrev, *Inyoites oweni* Hyatt et Smith, *Subvishnuites welteri* Spath, *Parussuria compressa* (Hyatt et Smith), *Lanceolites compactus* Hyatt et Smith, *L. bicarinatus* Smith, *Prosphingitoides austini* (Hyatt et Smith), *Owenites koeneni* Hyatt et Smith, and *Subowenites slavini* Popow. The taxonomic composition of the assemblage shows that the *Owenites-Dieneroceras* Beds are correlative with the *gracilitatis* chronozone of the middle Scythian (Fig. 2). Their

analogues are known in both the Tethyan and Boreal provinces (Kummel and Steele, 1962; Shevyrev, 1990). The above ammonoid assemblage of the northern Caucasus is most similar to that characterizing the *Meekoceras* Beds of the western United States (Hyatt and Smith, 1905; Smith, 1932).

The second ammonoid assemblage, distinguished in the Malyi Tkhach Formation, is relatively poor. It is characteristic of the *Stenopopanoceras* Beds and includes only seven species: *Stenopopanoceras transiens* Tozer, *Groenlandites? glaber* Shevyrev, *Longobarditoides caucasius* Shevyrev, *Laboceras gracile* Shevyrev, *Megaphyllites* sp., *M. compressus* Shevyrev, and *Leiophyllites pitamaha* (Diener). Among these species, only two (*S. transiens* and *L. pitamaha*) occur outside the northern Caucasus. The first of them is typical only of the lower Anisian (Tozer, 1972; Bucher, 1989); the latter form, which dominantly occurs at the same level, has also been reported from middle Anisian deposits. However, as *Stenopopanoceras* and *Groenlandites* are early Anisian genera as well, we may definitely suggest that the *Stenopopanoceras* Beds and the Malyi Tkhach Formation, respectively, represent a lower Anisian unit (Table 1). This conclusion contradicts the opinion of Yaroshenko (1978), who inferred an early Triassic age for this formation on the basis of palynological studies.

The four following ammonoid assemblages are confined to the Acheshbok Formation. They characterize the *Laboceras-Megaphyllites* and *Isculites* beds of its lower subformation and the *Phyllocladiscites-Nicomedites* and *Ptychites-Flexoptychites* beds of the upper subformation.

The *Laboceras-Megaphyllites* Beds are the richest in ammonoids. This level yielded the following 28 species of 20 genera: *Alanites visendus* Shevyrev, *Parasageceras tkhachense* Shevyrev, *Sgeceras* sp., *Norites labensis* Shevyrev, *Sturia sansovinii* (Mojsisovics), *Procladiscites elegans* Shevyrev, *Mesocladiscites caucasius* Shevyrev, *Caucasochordiceras simplex* (Shevyrev), *Pseudohollandites densicostatus* Shevyrev, *Robinsonites caucasius* Shevyrev, *Paradanubites palmatus* Shevyrev, *P. orbiculatus* Shevyrev, *P. inornatus* Shevyrev, *Laboceras gracile* Shevyrev, *L. acutulum* Shevyrev, *Longobarditoides caucasius* Shevyrev, *Aegeiceras byzovae* Shevyrev, *Japonites labaensis* Robinson, *Caucasites evolutus* Shevyrev, *C. inflatus* Shevyrev, *Gymnites* sp., *Megaphyllites prometheus* Shevyrev, *M. compressus* Shevyrev, *Leiophyllites pitamaha* (Diener), *L. visendus* Shevyrev, *L. ellipticus* Shevyrev, *Monophyllites kieperti* Toula, and *M. wengensis* (Klipstein). Four species of this assemblage (*L. gracile, L. caucasius, M. compressus*, and *L. pitamaha*) are inherited from the underlying *Stenopopanoceras* Beds. The biostratigraphic level of the assemblage is inferable from its generic composition. Outside the Caucasus, such genera as *Parasageceras, Paradanubites, Aegeiceras*, and *Caucasites* occur mostly in lower Anisian deposits. The genera *Alanites* and *Pseudohollandites* are confined everywhere to basal beds of the middle Anisian (the *osmani-ismidicus* chronozone). The form *M. kieperti* has been reported from the top part of the *ismidicus* Zone in Turkey (Fantini Sestini, 1988) and *Gymnites* Beds (*balatonicus* chronozone) of Iran. The taxon *Robinsonites* is similar to the genus *Unionvillites* from the *hyatti* Zone in Nevada (Silberling and Nichols, 1982), the interval of which corresponds to the *osmani-ismidicus* chronozone. Other ammonoids have wider stratigraphic ranges.

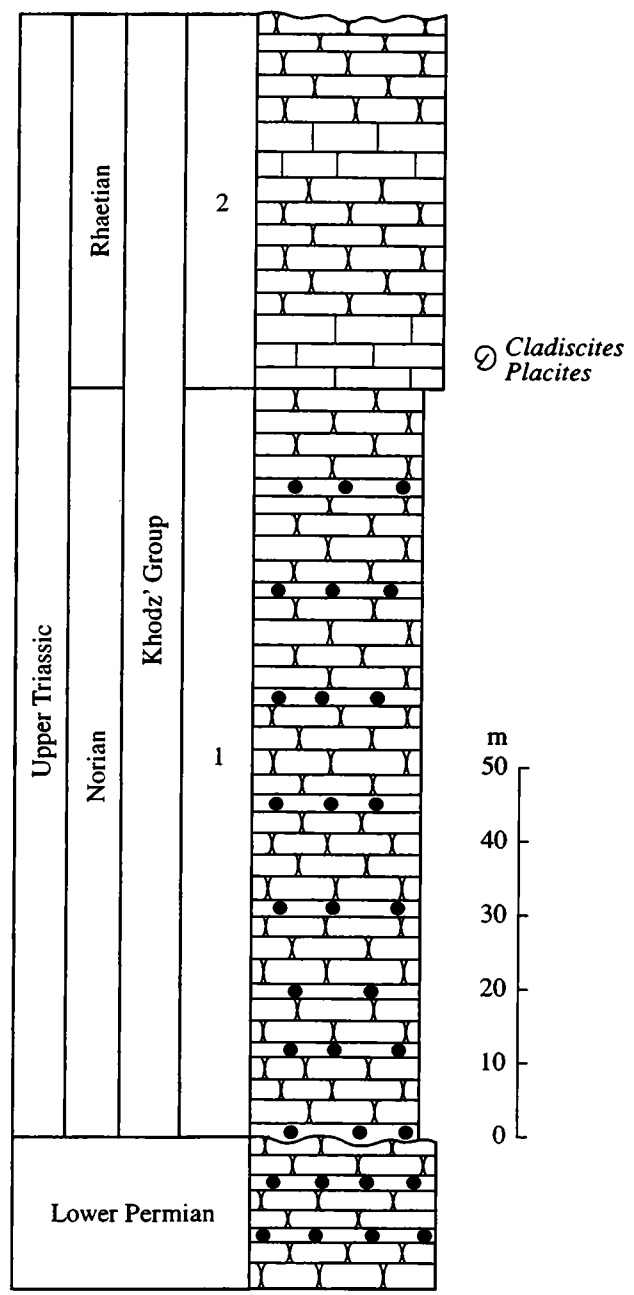

Fig. 6. Section of the Khodz' Group, Guzeripl' locality of the Belaya River (the author's interpretation based on data of Rostovtsev, 1973; symbols as in Fig. 2).

In general, these features clearly suggest that the *Laboceras-Megaphyllites* Beds correspond to the lower *osmani* chronozone of the middle Anisian (Table 1).

The dominant form of the *Isculites* Beds is *I. sphaericus* (Shevyrev), which is restricted in distribution to this particular level. Associated species are represented by *Sturia sansovinii, Megaphyllites compressus*, and *Leiophyllites visendus*, all transited at this level from the *Laboceras-Megaphyllites* Beds. The genus *Isculites* is usually confined to the basal level of the middle Anisian (the *osmani-ismidicus* chronozone) and occasionally to the lower Anisian. Because of the close relation of this assemblage with ammonoids typical of the *Laboceras-Megaphyllites* Beds, I tend to correlate the *Isculites* Beds with the *osmani* Zone.

The rich ammonoid assemblage from the *Phyllocladiscites-Nicomedites* Beds comprises the following 21 species of 15 genera: *Vorobyevites unicus* Shevyrev, *Sturia sansovinii* (Mojsisovics), *Phyllosphaerites obscurus* Shevyrev, *Procladiscites elegans* Shevyrev, *Phyllocladiscites proponticus* (Toula), *Psilocladiscites* sp., *Acrochordiceras subrotundum* Shevyrev, *Epacrochordiceras inflatum* Shevyrev, *E. compressum* Shevyrev, *Nicomedites caucasius* (Shevyrev), *N. tkhachensis* (Shevyrev), *Kocaelia lenticularis* Shevyrev, *Longobarditoides magnus* Shevyrev, *Gymnites robinsoni* Shevyrev, *G. evolutus* Shevyrev, *Gymnites* sp. 1, *Gymnites* sp. 2, *Proarcestes bramantei* (Mojsisovics), *Leiophyllites visendus* Shevyrev, *Monophyllites kieperti* Toula, and *M. caucasius* Shevyrev. These beds represent the first occurrence level for nearly all of the aforementioned forms, except for *S. sansovinii, L. visendus* and *M. kieperti*. The taxa *Acrochordiceras, Epacrochordiceras, Nicomedites*, and *Kocaelia* are typical genera of the middle Anisian. The first two taxa are restricted in distribution by the *ismidicus* and *balatonicus* zones, and the other two, by the *osmani* and *ismidicus* zones. The form *Ph. proponticus* is known from the *osmani-trinodosus* Zone of Turkey (Fantini Sestini, 1988) and from the Nazarkardeh Formation (the *osmani-ismidicus* Zone) of Iran (Krystyn and Tatzreiter, 1991). Specimens of *P. bramantei* occur in the *ismidicus-trinodosus* Zone of the eastern Alps, Yugoslavia, and Turkey. Species of *M. kieperti* are typical of both the *ismidicus* Zone in Turkey and the *Gymnites* Beds (the *balatonicus* Zone) in Iran. According to the presented data, the *Phyllocladiscites-Nicomedites* Beds are correlative, in my opinion, with the *ismidicus* Zone (Table 1).

The *Ptychites-Flexoptychites* Beds contain such forms as *Ptychites besnosovi* Shevyrev, *Flexoptychites bugunzhensis* Shevyrev, *Epacrochordiceras* sp., *Sturia* sp., and *Gymnites* sp. The *Ptychites* forms have stratigraphic ranges from the middle Anisian to the Ladinian, while those of the *Flexoptychites* genus occur in the interval comprising the uppermost middle (the *balatonius* Zone) and upper Anisian. The distribution of the *Epacrochordiceras* genus is restricted by the *ismidicus* and *balatonicus* zones. With due regard for their stratigraphic position, the *Ptychites-Flexoptychites* Beds of the Caucasus should be attributed to the *balatonicus* Zone of the middle Anisian (Table 1).

Formerly, ammonoids from the *Laboceras-Megaphyllites, Isculites, Phyllocladiscites-Nicomedites*, and *Ptychites-Flexoptychites* beds were considered to be a joint late Anisian assemblage of the *trinodosus* Zone (Shevyrev, 1968). The above data show that this viewpoint was misleading.

The *Bugunzhites-Parasturia* Beds represent the next stratigraphic level. They have been distinguished only within the basal part of the Babuk Formation exposed in the Bugunzha headstream area. The beds contain *Bugunzhites dagysi* Shevyrev, *Parasturia acutata* Shevyrev, *Sturia semiarata* Mojsisovics, and *Monophyllites wengensis* (Klipstein). Outside the Caucasus, representatives of the *Bugunzhites* genus are known from the Doab Beds of Afghanistan (Furon and Rosset, 1951) and from the *Daonella* Beds of Iran (Krystyn and Tatzreiter, 1991). These beds are of the late Ladinian age. In the eastern Alps, the *Parasturia* forms and *S. semiarata* are typical of the *archelaus* Zone of the upper Ladinian, while the stratigraphic range of *M. wengensis* spans between the middle Anisian and upper Ladinian. Therefore, I consider the *Bugunzhites-Parasturia* Beds as chronologically corresponding to the *archelaus* Zone (Table 1).

Only *Proarcestes planus* Shevyrev and *Phloioceras* sp. have been identified in rocks of the Babuk Formation exposed along the Tkhach River. Ammonoids of the *Proarcestes* genus existed within the middle Anisian-late Carnian interval and were most prosperous at the end of this period. Nautiloids of the *Phloioceras* genus are known from the Ladinian-Norian deposits, and the time of their prosperity is also confined to the Carnian (Schastlivtseva, 1988). Thus, we may conventionally presume that the *Proarcestes-Phloioceras* Beds have a Carnian age.

Ammonoids from the *Goniojuvavites-Pararcestes* Beds of the Babuk Formation are more diverse. Their species are represented by *Clionitites* sp., *Projuvavites caucasius* Shevyrev, *Goniojuvavites planus* Shevyrev, *Proarcestes* sp., *P. planus* Shevyrev, *P. wittenburgi* Shevyrev, *Pararcestes acutus* (Mojsisovics), *P. optimus* Kurgalimova, *P. paropertus* Kurgalimova, and *P. sublabiatus* Mojsisovics. The *Clionitites* sp., *P. acutus*, and *P. sublabiatus* are the index species of the Carnian, and forms of the *Projuvavites* and *Goniojuvavites* genera mark the upper Carnian and, sometimes, the lowermost Norian deposits. Thus, the *Goniojuvavites-Pararcestes* Beds seem to correspond to the upper Carnian interval (Table 1).

The *Placites-Rhacophyllites* Beds of the Khodz' Formation yielded the last ammonoid assemblage of the Triassic sequence in the northern Caucasus. It consists of the following forms: *Cladiscites tornatus* (Bronn), *Paracladiscites multilobatus* (Bronn), *Ady-*

Table 1. Correlation of the Middle-Upper Triassic deposits in the northern Caucasus with the standard ammonoid zones

Standard scale					Northern Caucasus		
Upper Triassic	Rhaetian	Upper	*Choristoceras marshi*	*Choristoceras marshi*	Khodz' Group	Khodz' Formation	
				Vandaites stuerzenbaumi			*Placites-Rhacophyllites* Beds
			Rhabdoceras suessi	*Sagenites reticulatus*			
		Lower		*Sagenites quinquepunctatus*		Shapka Formation	*Monotis* Beds
	Norian	Upper	*Halorites macer*				
			Himavatites hogarti				
			Cyrtopleurites bicrenatus				
		Lower	*Juvavites magnus*				
			Malayites paulckei				
			Guembelites jandianus				
	Carnian	Upper	*Anatropites spinosus*		Sakhrai Group	Babuk Formation	*Goniojuvavites-Pararcestes* Beds
			Tropites subbullatus				
			Tropites dilleri				
		Lower	*Austrotrachyceras austriacum*				*Proarcestes-Phloioceras* Beds
			Trachyceras aonoides				

Table 1. (Contd.)

Standard scale				Northern Caucasus		
Middle Triassic	Ladinian	Lower	*Frankites regoledanus*			
			Protrachyceras archelaus			*Bugunzhites-Parasturia* Beds
		Upper	*Eoprotrachyceras curionii*			
	Anisian	Lower	*Xenoprotrachyceras reitzi*			
			Aplococeras avisianum			
			Paraceratites trinodosus			
		Upper	*Balatonites balatonicus*	Tkhach Group	Acheshbok Formation	*Ptychites-Flexoptychites* Beds
			Aghdarbandites ismidicus			*Phyllocladiscites-Nicomedites* Beds
			Nicomedites osmani			*Isculites* Beds
						Laboceras-Megaphyllites Beds
		Lower	*Aegeiceras ugra*		Malyi Tkhach Formation	*Stenopopanoceras* Beds

geites obscurus Shevyrev, *Placites polydactylus* (Mojsisovics), *Pinacoceras metternichi* (Hauer), *Megaphyllites insectus* (Mojsisovics), *Stenarcestes leiostracus* (Mojsisovics), and *Rhacophyllites debilis* (Hauer). Such forms as *P. multilobatus, P. metternichi*, and *Rh. debilis* occur in both the Norian and Rhaetian deposits, while *C. tornatus, P. polydactylus, M. insectus*, and *S. leiostracus* are typical of the Rhaetian only, mostly of its *suessi* Zone and basal horizons (the *stuerzenbaumi* Subzone) of the *marshi* Zone. In addition, it is remarkable that the beds under consideration occupy a considerably higher level than the *Monotis* Beds, which are the stratigraphic equivalent of the *quinquepunctatus* Subzone. They are also above the boundary that marks significant changes in the taxonomic composition of brachiopods between members 6 and 7 of the Khodz' Formation exposed on Mt. Yatyrgvarta (Fig. 5). Assuming that this level corresponds to the lower boundary of the *marshi* Zone, I refer the *Placites-Rhacophyllites* Beds to the lower part of this zone (the *stuerzenbaumi* Subzone; Table 1).

One of remarkable features characterizing the general section of Triassic deposits in the northern Caucasus is represented by two hiatuses. One of them separates the Tkhach and Sakhrai groups. It corresponds to the interval of either the upper Anisian and entire Ladinian (Fig. 3) or the upper Anisian and lower Ladinian substages (Fig. 4; Table 1). Formerly, the lack of Ladinian deposits in the Tkhach basin was recognized by Yaroshenko (1978), who studied miospores from this region. The second hiatus separates the Sakhrai and Khodz' groups. It is difficult to evaluate its duration,

because the lower part of the Shapka Formation, the basal one in the Khodz' Group, yielded very poor paleontological remains. Presumably, this hiatus corresponds to the lower Norian.

Another significant feature of Triassic deposits in the northern Caucasus is manifested in the abundance and diversity of middle Anisian ammonoids, the succession of which is surprisingly regular. In their taxonomic composition, these ammonoid assemblages are most similar to their analogues from Turkey and Iran, which eliminates many problems in correlating the respective host rocks.

One of the well-known Anisian sections is in Turkey, near the Gebze on the northern coast of the Gulf of Ismid (Assereto, 1972, 1974; Fantini Sestini, 1988). The cloddy limestones (114 m) exposed here are divisible into three members with eight litho-stratigraphic levels (Table 2). The lower member (34 m) comprises four levels (1–4), all barren of fossils and conventionally attributed to the lower Anisian (the Aegean). The middle member (31 m), containing three levels (5–7), corresponds to the lower part of the middle Anisian (the Bithynian). Its lower level (11 m) is known as the *Nicomedites osmani* Zone, which comprises the *Grambergia? solimani* and *Kocaelia toulai* subzones. The two other levels (12 and 8 m) span the *Aghdarbandites ismidicus* Zone. The lower unit represents the first occurrence level of the genus *Acrochordiceras*, which have shells covered with large tubercles (similar to *A. asseretoi*). The upper unit is the occurrence level of *Aghdarbandites ismidicus*. The main part of the upper member (49 m, level 8) represents the *Balatonites balatonicus* Zone of the Pelsonian with the *Epacrochordiceras pustericum* and *Balatonites balatonicus* subzones. As for the uppermost portion of this member, Italian researchers believe that, near Gebze, it corresponds to the *Paraceratites trinodosus* Zone (the Illyrian) with the *Paraceratites binodosus* and *Flexoptychites flexuosus* subzones. However, *P. binodosus* was found in association with *B. balatonicus* (Fantini Sestini, 1988: Figs. 7 and 10), and it seems unreasonable to separate the *binodosus* and *balatonicus* zonal units and to include the former into the *trinodosus* chronozone. The terminal Anisian beds above the last level consist of red limestones of the "ammonitico rosso" facies. The suggested correlation of ammonoid-bearing deposits of the northern Caucasus with the described section of Turkey is shown in Table 2.

In Iran, there are two interesting outcrops. One of them is in the northeastern region of the country near Aghdarband, a settlement located southeast of Mashhad (Krystyn and Tatzreiter, 1991). Anisian ammonoids have been recovered here from the Nazarkardeh Formation (up to 50 m), which consists of cloddy, partially silicified limestones alternating with shales and sandstones. Despite nonsystematic sampling, ammonoids from these deposits are easily divisible into two assemblages. One of them includes the following forms: *Nicomedites osmani* Toula, *Pseudohollandites eurasiaticus* Krystyn et Tatzreiter, *Gymnites aghdarbandensis* Krystyn et Tatzreiter, *Costogymnites asiaticus* Krystyn et Tatzreiter, *Procladiscites* cf. *proponticus* Toula, and *Leiophyllites suessi* (Mojsisovics). The other assemblage is represented by *Aghdarbandites ismidicus* (Arthaber), *Semibeyrichites ruttneri* Krystyn et Tatzreiter, *Gymnites asseretoi* Tozer, *"Japonites"* cf. *kirata* Diener, *Sturia sansovinii* (Mojsisovics), *Procladiscites* cf. *proponticus* Toula, and *Psilosturia* sp. These assemblages respectively mark the *osmani* and *ismidicus* zones of the middle Anisian (Table 2). The Nazarkardeh Formation conformably rests on the Safid Kuh Limestone, presumably representing the lower Anisian unit. Above it, there are tuffaceous sandstones and shales of the Sina Formation bearing bivalves and ammonoids of the upper Ladinian. Thus, similar to deposits in the northern Caucasus, the Aghdarband section displays a hiatus spanning the upper Anisian and lower Ladinian.

The outcrops of central Iran have been studied at Mt. Nakhlak, southeast of the Anarak (Davoudzadeh and Seyed-Emami, 1972; Tozer, 1972). The Triassic section begins here with the Alam Formation (870 m), comprising six members of pure or shaly limestones. They are subdivided into nine ammonoid beds, four of which are the late Scythian in age and, thus, outside the scope of this work. The Anisian corresponds to five other beds with the following characteristics (Table 2): the *Ussurites-Stenopopanoceras* Beds with *Ussurites arthaberi* (Welter), *Stenopopanoceras transiens* Tozer, and *Procladiscites* sp. represent the lower Anisian; the *Leiophyllites* Beds with *Norites gondola* (Mojsisovics), *Hollandites tozeri* Zakharov, *"Hungarites"* cf. *proponticus* Toula, and *Leiophyllites stoecklini* Tozer span the *osmani* Zone; the *Anagymnotoceras* Beds with *Nicomedites* cf. *toulai* (Arthaber), *"Anagymnotoceras" ismidicum* (Atrhaber), and *Gymnites religiosus* Diener are equivalent to the *ismidicus* Zone; and both the *Epacrochordiceras* Beds with *Epacrochordiceras* sp. and the *Gymnites* Beds with *Acrochordiceras* aff. *hyatti* Meek, *Paraceratites* aff. *binodosus* (Hauer), *Gymnites palmai* (Mojsisovics), *G. asseretoi* Tozer, *Parapinacoceras* cf. *damesi* (Mojsisovics), *Sturia* sp., *Ptychites* aff. *pauli* Mojsisovics, *Leiophyllites* aff. *pitamaha* (Diener), and *Monophyllites kieperti* Toula correspond to the *balatonicus* Zone. The upper member of the Alam Formation and the overlying Baqoroq Formation are barren of fossils. According to their stratigraphic position, they can be correlated with the upper Anisian–lower Ladinian interval. Higher in this section, ammonoids appear only in the *Daonella* Beds of the upper Ladinian Ashin Formation.

The Anisian ammonoids from the northern Caucasus are also similar to their counterparts in China. A remarkable Chinese section is located on the southern slope of Mt. Burhan Budai, central Qinghai (He *et al.*, 1986). Here, the Naocangjianguo Formation, with abundant lower-middle Anisian ammonoids, rests

Table 2. Interregional correlation of Anisian deposits

Standard	Turkey, Gebze	Northern Caucasus	Iran: Aghdarband	Iran: Anarak	China, Qinghai	Nevada, USA
Upper Anisian	Illyrian					
reitzi	Uppermost Anisian					
avisianum						
trinodosus	*Flexoptychites flexuosus*					
	Redlimestones of ammonitico rosso facies					
				Baqoroq Formation		
						Frechites occidentalis
						Parafrechites meeki
						Gymnotoceras rotelliforme

Table 2. (Contd.)

Standard	Middle Anisian							
	osmani		*ismidicus*		*balatonicus*			
Turkey, Gebze	*osmani*	*Kocaelia toulai*	*Acrochordiceras asseretoi*	*Aghdarbandites ismidicus*	*Epacrochordiceras pustericum*	*Balatonites balatonicus*	*Paraceratites trinodosus*	*trinodosus*
	Bithynian				Pelsonian			
	5	6	7	Middle member	8	Upper member		Cloddy limestones
Northern Caucasus	*Laboceras-Megaphyllites* Beds	*Isculites* Beds	*Phylloceltidiscites-Nicomedites* Beds		*Prychites-Flexoprychites* Beds			Achesbok Formation
	Tkhach Group							
Iran, Aghdarband	Nazarkardeh Formation							
Iran, Anarak	*Leiophyllites* Beds		*Anagymnoceras* Beds		*Epacrochordiceras* Beds		*Gymnites* Beds	Alam Formation
	Nakhlak Group							
China, Qinghai	*Gymnites*		*Nicomedites-Anagymnoceras*		*Semiornites* Beds			Naocangjiangou Formation
	Beyrichites-Hollandites							
Nevada, USA	Lower		Upper					Prida (Favret) Formation
	Acrochordiceras hyatti				*Nevadisculites taylori*		*Balatonites shoshonensis*	Middle (Fossil Hill) member

Table 2. (Contd.)

Standard	Turkey, Gebze	Northern Caucasus	Iran: Aghdarband	Iran: Anarak	China, Qinghai	Nevada, USA
Lower Anisian						
ugra						
	Aegean					
	ugra					
	Lower member					
	1–4					
		Malyi Tkhach Formation				
		Stenopopanoceras Beds				
			Sefid Kuh Limestone			
				Ussurites-Stenopopanoceras Beds		
					Lenotropites-Japonites	
					Lenotropites qinghaiensis	*Japonites meridianus*
Japonites welteri Beds	*Pseudo-keyserlingites guexi* Beds				*Silberlin-gites mulleri*	*Lenotropites caurus*

on graywakes and sandstones (the Hongshuichuan Group) identified as the *Subcolumbites* Beds of the upper Scythian. The basal marl member (318 m) of this formation, corresponding to the *Lenotropites-Japonites* Zone of the lower Anisian, yielded 52 species of 18 ammonoid genera. The zone comprises the *Lenotropites qinghaiensis* and *Japonites meridianus* subzones (Table 2). Longobarditids, represented by forms of *Grambergia, Lenotropites, Groenlandites*, and *Tienjunites* genera, are dominant in the lower subzone, where they associate with species of *Arctohungarites, Paradanubites, Keyserlingites, Norites, Procladiscites, Megaphyllites*, and *Leiophyllites* genera. Among ammonoids of the upper subzone, the most abundant specimens belong to *Japonites, Psilosturia, Paradanubites, Procladiscites, Keyserlingites, Megaphyllites*, and *Ussurites* genera. Other taxa encountered at this level include *Parasageceras, Sageceras, Psilocladiscites*, and *Metajaponites* genera. The next member of the Naocangjianguo Formation consists of sandstones and siltstones alternating with conglomerates and limestones. It represents the *Beyrichites-Hollandites* Zone, where 25 species of 14 ammonoid genera have been distinguished. The zone includes the *Gymnites* and *Nicomedites-Anagymnotoceras* subzones, the lower of which is dominated by specimens of *Gymnites, Buddhaites, Psilosturia*, and *Ussurites* genera. In addition, this level yielded the first beyrichitids (genera *Beyrichites* and *Philippites*), and species of *Longobarditoides* and *Isculites* genera. Beyrichitids (genera *Beyrichites, Nicomedites*, and *Hollandites*) become dominant in the upper subzone, where they occur together with abundant representatives of *Anagymnotoceras, Sturia, Longobarditoides, Procladiscites*, and *Leiophyllites* genera. The lower of these units corresponds to the *osmani* chronozone, and the upper one, to the *ismidicus* Zone. According to Chinese scientists, the *Beyrichites-Hollandites* Zone underlies the *Semiornites* Beds, which are the equivalent of the *balatonicus* and *trinodosus* chronozones.

The richest assemblages of Anisian ammonoids are known from northwestern Nevada (USA), where the Prida and Favret formations represent two lower interrelated units of the Star Peak Group (Nichols and Silberling, 1977).

The Prida Formation, exposed in canyons of Humboldt Range, consists of three members (Silberling and Wallace, 1969). The lower member includes basal conglomerate and sandstone (15 m), limestone with shale interlayers (100 m), and brown limy sandstone (30 m). The upper portion of the limestone yielded ammonoids of the *Neopopanoceras haugi* Zone. Formerly, this zone was attributed to the Anisian (Hyatt and Smith, 1905; Spath, 1934; Kummel, 1957); however, recently, many researchers have interpreted it as corresponding to the uppermost Scythian (Silberling and Tozer, 1968; Silberling and Wallace, 1969; Bucher, 1989).

The middle Fossil Hill Member (180 m) of the Prida Formation comprises sandstones, clays, and shales. In the basal one-third of the member, the rocks are subdivided into the *Japonites welteri* and *Pseudokeyserlingites guexi* beds of the *Silberlingites mulleri* and *Lenotropites caurus* zones. These units correspond to the lower Anisian (Table 2), as they yielded 27 species of such genera as *Japonites, Hemilecanites, Metadagnoceras, Paracrochordiceras, Pseudokeyserlingites, Leiophyllites, Grambergia, Ussurites, Caucasites, Silberlingites, Paradanubites, Sageceras, Groenlandites, Gymnites, Isculites*, and *Lenotropites* (Bucher, 1989).

At a level of 40 m and higher in the section, there is the *Acrochordiceras hyatti* Zone. Ammonoids from this interval are represented by 20 species of the following genera: *Koipatoceras, Alanites, Ismidites, Amphipopanoceras, Acrochordiceras, Cuccoceras, Nicomedites, Hollandites, Anagymnotoceras, Japonites, Gymnites, Isculites, Czekanowskites, Pseudodanubites, Unionvillites, Intornites*, and *Ussurites* (Silberling and Nichols, 1982). The zone has been subdivided into lower and upper subzones that correspond to the *osmani* and *ismidicus* chronozones, respectively.

Next, the *Nevadisculites taylori* Zone, best studied in the Augusta Range, was also recently recognized within the Fossil Hill Member of the Favret Formation as overlying the *hyatti* Zone (Bucher, 1988). It yielded 25 species representing genera *Pseudodanubites, Platycuccoceras, Anagymnotoceras, Ptychites, Augustaceras, Epigymnites, Eogymnotoceras, Nicomedites, Ismidites, Nevadisculites, Megaphyllites, Acrochordiceras, Epacrochordiceras, Intornites, Ussurites*, and *Sageceras*. The zone is divisible into four units: the *Pseudodanubites nicholsae, Anagymnotoceras* cf. *spivaki, Augustaceras escheri*, and *Platycuccoceras praebalatonense* beds. Above the *taylori* Zone of the Augusta Range localities, there is the *Balatonites shoshonensis* Zone. This level of Triassic deposits in Nevada contains 26 species of such genera as *Eogymnotoceras, Gymnotoceras, Ismidites, Constrigymnites, Nevadisculites, Ussurites, Epacrochordiceras, Intornites, Acrochordiceras, Ptychites, Favreticeras, Amphipopanoceras, Balatonites, Platycuccoceras, Guexites, Proteusites, Proarcestes*, and *Bulogites*. The present *shoshonensis* Zone is subdivided into four units: the *Favreticeras rieberi, F. ransomei, F. wallacei*, and *Proteusites fergusoni* subzones (Bucher, 1992). The most remarkable ammonoids of this zone are representatives of the *Balatonites* genus, which suggest that it is correlative with the *balatonicus* chronozone of the middle Anisian (Table 2). Its partial equivalents are known in the eastern Alps (Assereto 1971; Balini, 1993), Hungary (Vörös, 1987; Tatzreiter and Vörös, 1991), Greece (Krystyn and Mariolakos, 1975), Turkey (Fantini Sestini, 1988), central Iran (Tozer, 1972), and northern Tibet (Gu *et al.*, 1980). It is more difficult to find correlatives for the *taylori* Zone of North America. Thus far, I have conventionally considered it as corresponding to the basal portion of the *balatonicus* Zone.

The uppermost beds of the Fossil Hill Member of the southern Humboldt Range represent units that are very rich in late Anisian ammonoids that have been known since the early 1900s (Hyatt and Smith, 1905; Smith, 1914). Revising this diverse assemblage, Silberling and Nichols (1982) identified 37 species of 18 genera, and they distinguished 12 ammonoid beds in the respective interval of the section, which is only 25–30 m thick. The five lower beds correspond to the *Gymnotoceras rotelliforme* Zone. The three following beds are correlative with the *Parafrechites meeki* Zone, and the four upper beds represent the *Frechites occidentalis* Zone. These local zones correspond to the *trinodosus, avisianum*, and *reitzi* chronozones, respectively. There are no equivalents of these zones in the northern Caucasus (Table 2).

The upper member of the Prida Formation is the Ladinian in age. In this unit, ammonoids are less abundant than in the Fossil Hill Member, especially in its upper part. Basal beds (10 m) of the member represent the stratotype of the local *Protrachyceras subasperum* Zone, which is correlative with the *curionii* chronozone of the lower Ladinian. In the northern Caucasus, corresponding deposits are also unknown.

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Stratigraphy and Lithology of the Upper Triassic Deposits of Southwestern Cyprus (Vlambouros Formation)

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Abstract—The stratigraphic interval of terrigenous deposits of the Vlambouros Formation is determined as Upper Carnian–Norian by radiolarians and conodonts. A large stratigraphic hiatus, which includes the Lower Jurassic and main part of the Middle Jurassic, is distinguished between the Vlambouros Formation and overlying Episkopi Formation (Upper Callovian–Cretaceous). The deposition of the Vlambouros Formation is interpreted as the result of erosion of the continental supracrustal assemblage and adjacent carbonate shelf with subsequent turbidite and contourite sedimentation. The neritic character of the Late Norian radiolarian assemblage is assumed.

Key words: *Cyprus, Tethys, Mamonia, passive margin, Triassic, Jurassic, stratigraphy, hiatus, Vlambouros, Episkopi, turbidites, contourites, currents.*

The allochtonous Mamonia Complex attracts particular attention among the various geologic units of southwestern Cyprus. It includes the most ancient deposits of this region. Their study is of primary importance for understanding the Early Mesozoic history of Cyprus and other regions of the eastern Mediterranean. Previous researchers subdivided the Triassic deposits of the Mamonia Complex into two units: the volcanogenic, with lenses and intercalations of sedimentary rocks, and completely sedimentary (Lapierre, 1975). This sedimentary unit was identified later as the Vlambouros Formation of presumably Middle to Late Triassic age (Robertson and Woodcock, 1979).

Different sandstones, usually with plant detritus, and siltstones are the predominant rocks of the Vlambouros Formation. Calcarenites and micritic limestones with remains of *Halobia* occur as intercalations. The Marona Formation (Fig. 1), represented by monotonous micritic limestones of the Late Triassic age, was distinguished in some sections above the Vlambouros Formation (Robertson and Woodcock, 1979). The Triassic deposits are overlain by the cherty unit with basal red siltstones supposed to be Early Jurassic in age (Robertson and Woodcock, 1979). Some authors think that the uppermost part of the Vlambouros Formation corresponds to the Early Jurassic, and overlying cherts to the Dogger–Malm (Cleintuar *et al.*, 1974). In any case, Triassic clastic deposits of the Vlambouros Formation and overlying cherts and clastics of the Episkopi Formation were described as a complete sequence spanning the Upper Triassic–Middle Cretaceous interval (Robertson and Woodcock, 1979; Swarbrick and Robertson, 1990). The Upper Triassic Vlambouros Formation was interpreted as sediments of the southern passive margin of Tethys deposited at the rifting stage (Kaz'min *et al.*, 1987; Robertson and Woodcock, 1979). Permanent deep-water conditions of sedimentation during the Jurassic and Early Cretaceous presumably appeared after this rifting (Robertson, 1990). The similar structure of the sequence is typical of adjacent areas, northwestern Syria (Delaune-Mayere *et al.*, 1977) and southwestern Turkey (Robertson, 1990; Robertson and Woodcock, 1981).

The detailed biostratigraphy and lithology of the Triassic deposits of the Vlambouros Formation was studied in 1990 during the field work of the expedition of the Geological Institute of the Academy of Sciences of the USSR. Numerous new data, which were obtained during this study, allowed us to revise many previous conclusions on the stratigraphy and geological history of the region. We studied main sections in the area of distribution of the Vlambouros Formation (Fig. 2). The most complete sequence is located near the spring Argakin ton Vatkion on the left bank of the Mavrokolymbos River. This sequence yielded the best paleontological remains. It is represented by the following units:

1. Yellowish gray, coarse- to middle-grained, thick-bedded sandstones with well developed A–C Bouma intervals. The thickness of some beds is 1 m. Sandstones bear abundant plant detritus confined to the bedding surfaces and remains of *Halobia* sp. The observed thickness is 2 m.

2. Light gray, middle- to fine-grained, massive and thick-bedded sandstones with numerous intercalations of greenish gray siltstones and rare beds of light gray, micritic limestones. Sandstones have typical turbidity structures with distinctive B–C and B–C–D–E Bouma intervals. The limestones contain Carnian to early Norian conodonts *Neogondolella nodosa* (Hayashi). The thickness is 10 m.

System	Robertson and Woodcock, 1979	This study		System Series Stage
Jurassic	Episkopi Formation	Episkopi Formation		J_2cl-J_3ox
Triassic	Marona Formation	Vlambouros Formation	upper member	T_3n_3-r
			middle member	T_3n_3-r
	Vlambouros Formation		lower member	T_3k_3-n_2

Fig. 1. Stratigraphy of the Triassic deposits of southwestern Cyprus.

3. Greenish gray, fine-grained sandstones and siltstones with intercalations of gray and pink micritic limestones and pink calcarenites. Limestones bear the late Carnian to Norian conodonts *Epigondolella* sp. The thickness is 8 m.

4. Gray and reddish gray sandstones, intercalated with greenish-gray, or, sometimes, reddish gray, fine-grained siltstones, red cherts and pink calcarenites. Cherts contain the late Norian to Rhaetian (?) radiolarians *Livarella gifuensis* Yoshida, *L.* sp., and *Betraccium* sp. The thickness is 10 m.

5. Red, middle- to fine-grained sandstones with carbonate cement, intercalated with brick red siltstones. The thickness is 5 m.

6. Gray, fine-grained sandstones, intercalated with brick-red siltstones and rare beds of red cherts with the late Norian to Rhaetian (?) radiolarians *Livarella* sp. The thickness is 10 m.

These sediments are covered without visible discordance by red cherty mudstones with numerous thin (3 cm) intercalations of red and green cherts bearing the late Callovian to early Oxfordian radiolarians *Gueuxella nudata* (Kocher), *Hsuum brevicostatum* Ozwoldova, *Protunuma ochiensis* Matsuoka, and *Zhamoidellum mikamense* Aita. These sediments were considered as basal layers of the Episkopi Formation.

The next section was observed near the Lara Bay, the Akamas Peninsula; its units are as follows:

VLAMBOUROS FORMATION

1. Gray, yellowish and reddish gray, fine-grained, cross-bedded, platy sandstones with intercalations of grey, cross-bedded siltstones, pink calcarenites, grey and pink limestones and rare concretions of sandstone with carbonate cement. The observed thickness is 20 m.

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2. Red, brick red, sometimes light green, platy cherts with the late Callovian to early Oxfordian radiolarians: *Eucyrtidiellum ptyctum* (Riedel et Sanfilippo), *Gueuxella nudata* (Kocher), *Haliodictya hojnosi* (Riedel et Sanfilippo), *Higumastra imbricata* (Ozwoldova), *Hsuum brevicostatum* (Ozwoldova), *Mirifusus guadalupensis* (Pessagno), *Parvicingula dhimenaensis* (Baumgartner), *Protunuma ochiensis* (Matsuoka), *Ristola altissima* (Rust), and *Zhamoidellum mikamense* Aita. The thickness is 12 m.

Member 1 of this sequence can be correlated with members 3 and 4 of the sequence at the Mavrokolymbos River. The overlying sandstones, intercalated with

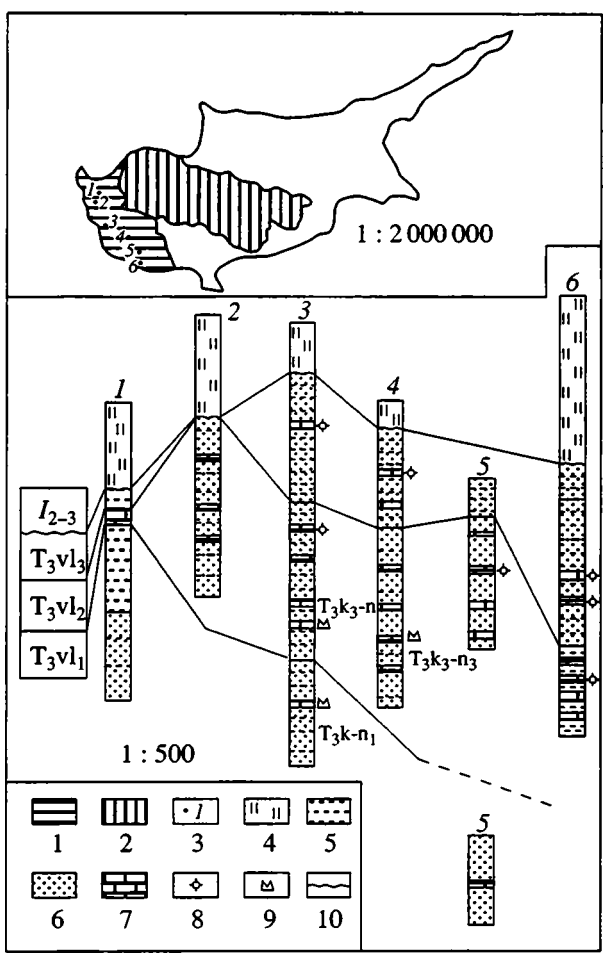

Fig. 2. Stratigraphic sections of the Vlambouros Formation and their correlation: (*1*) Phasli; (*2*) Lara; (*3*) Mavrokolimbos; (*4*) Episkopi; (*5*) Dhiarizos; (*6*) Khapotami.

1—area of Mamonia Complex, 2—troodos ophiolite massif, 3—numbers of sections and their position, 4—cherts. 5—mudstones and siltstones, 6—sandstones, 7—limestones and calcarenites, 8—locations of Late Norian radiolarians, 9—locations of conodonts, 10—stratigraphic discordance at the basal part of Episkopi Formation.

Plate I

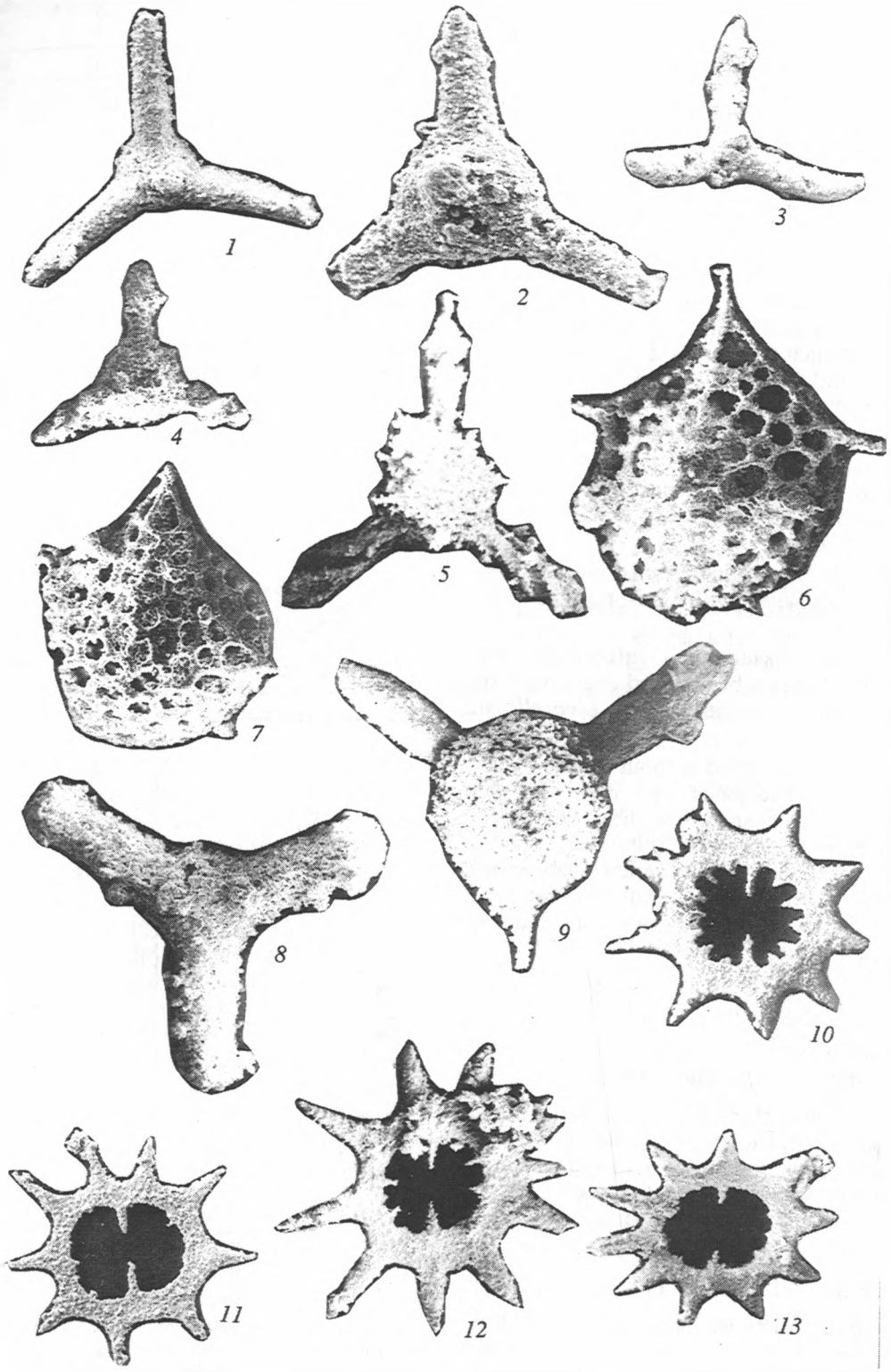

Plate I. Late Norian radiolarians from the section of the Dhiarizos River (member 2).

1—Livarella longus Yoshida; *2—Livarella* sp. cf. *L. densiporata* Kozur et Mostler; *3—Livarella gifuensis* Yoshida; *4, 5—Betraccium* sp. cf. *B. deweveri* Pessagno et Blome; *6—Pentactinocarpus magnus* (Kozur et Mostler); *7—Pentactinocarpus* sp. cf. *P. sevaticus* Kozur et Mostler; *8—Livarella* sp.; *9—Neopaurinella* sp. cf. *N. sevatica* Kozur et Mostler; *10—Praemesosaturnalis multidentatus* (Kozur et Mostler); *11—Praemesosaturnalis* sp. I; *12, 13—Praemesosaturnalis* sp. aff. *P. gracilis* (Kozur et Mostler).

All figures have magnification ×110.

Plate II

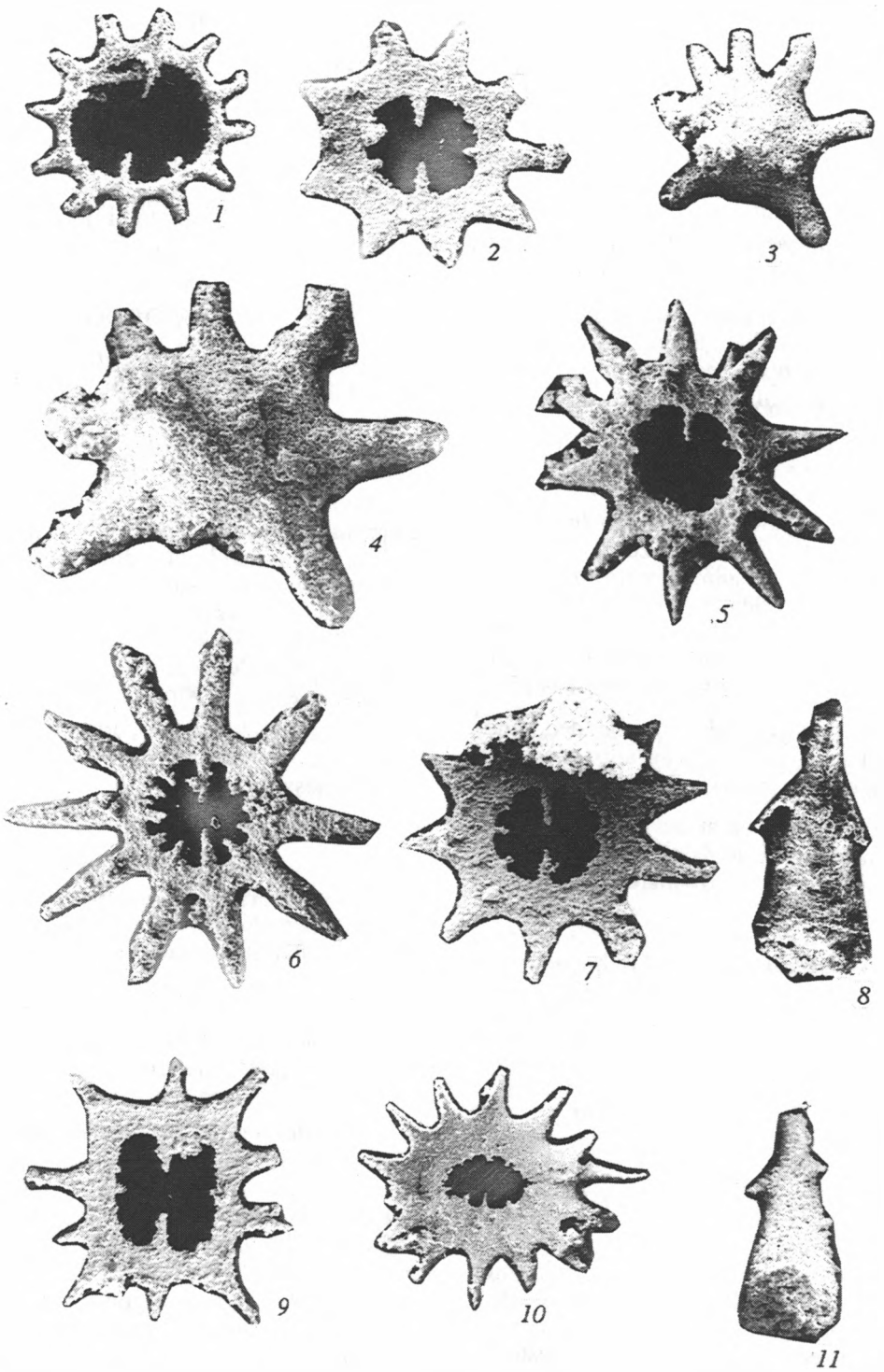

Plate II. Late Norian radiolarians from the section of the Dhiarizos River (member 2).
1—*Praemesosaturnalis zapfei* (Kozur et Mostler); *2*—*Praemesosaturnalis quadriradiatus* (Kozur et Mostler); *3, 4*—*Livarella*-like forms: *5*—*Praemesosaturnalis* sp. aff. *P. gracilis* (Kozur et Mostler); *6*—*Praemesosaturnalis* sp. II; *7*—*Praemesosaturnalis* sp.; *8, 11*—*Tetraspinocyrtis* (?) sp.; *9*—*Praemesosaturnalis* sp. III.; *10*—*Praemesosaturnalis* sp. aff. *P. multidentatus* (Kozur et Mostler). Magnification ×55 for specimen 3, and ×110 for other specimens.

siltstones (members 5 and 6 of the Mavrokolymbos River section), are absent here, and basal layers of the Episkopi Formation (upper Callovian–lower Oxfordian) are in direct contact with the middle part of the Vlambouros Formation. Unfortunately, there are no fossil finds from the sediments of the Vlambouros Formation at the Lara Bay. They can be correlated only lithologically with the members of other sections.

The strongly deformed sequence of the Vlambouros Formation can be partially seen at the right side of the Dhiarizos River opposite the village of Souskiou. This locality is in the central part of the Mamonia Complex area. The section includes the following beds:

1. Gray, middle- to fine-grained, thick-bedded sandstones with plant detritus and rare intercalations of gray micritic limestones. The observed thickness is 10 m below the unexposed interval of 30–40 m.

2. Gray and greenish gray, fine-grained sandstones with frequent intercalations of pink calcarenites and gray micritic limestones with the late Norian radiolarians: *Betraccium* sp. cf. *B. deweveri* Pessagno et Blome, *Livarella* sp. cf. *L. densiporata* Kozur et Mostler, *L. longus* Yoshida, *L.* sp., *Pentactinocarpus* sp. cf. *P. sevaticus* Kozur et Mostler, *P. magnus* (Kozur et Mostler), *Praemesosaturnalis* sp. aff. *P. gracilis* (Kozur et Mostler), *P. multidentatus* (Kozur et Mostler), and others (see Plates I and II). The observed thickness is 15 m.

3. Reddish gray, fine-grained, cross-bedded sandstones, intercalated with reddish gray siltstones with rare beds of red cherts. The observed thickness is 5 m.

A similar section is exposed at the right side of the Ezousa River (2 km downstream from the Episkopi Village). Here, it consists of the following units:

VLAMBOUROS FORMATION

1. Gray, medium-grained, massive sandstones with interlayers of greenish gray, middle- to fine-grained, sometimes silty, platy sandstones, which contain intercalations of greenish gray and light gray, micritic limestones. Massive sandstones contain intercalations of pink calcarenites and lenselike concretions of white and pink, micritic limestones with the late Carnian–late Norian conodonts *Epigondolella abneptis* (Huckriede). The observed thickness is 20 m.

2. Gray, middle-grained calcarenites (calciturbidites) and sandstones with carbonate cement and frequent intercalations of brown and brick red siltstones alternating with rare beds of gray micritic limestones and pink cherts, which yielded the late Norian–Rhaetian (?) radiolarians: *Betraccium* sp. and *Livarella densiporata* Kozur et Mostler. The thickness is 8 m.

3. Gray, fine-grained, silty, cross-bedded, gradational sandstones interbedded with reddish brown mudstones with thin interlayers of siltstones. The thickness is 3 m.

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4. Greenish, pink and brick red, sometimes banded cherts, intercalated with brick red, cherty mudstones. Cherts bear the Callovian–Oxfordian radiolarians: *Higumastra imbricata* (Ozwoldova), *Hsuum brevicostatum* (Ozwoldova), and *Protynuma ochiensis* Matsuoka. The thickness is 3.5 m.

The Vlambouros Formation is slightly different in composition in the Khapotami River basin (southeastern part of the studied area), where the sequence at the right side of the Khapotami River (3 km from its mouth) includes the following beds:

VLAMBOUROS FORMATION

1. Reddish gray and gray siltstones interbedded with gray, micritic limestones, pink, cross-bedded calcarenites and gray, fine-grained sandstones. Limestones bear the late Norian–Rhaetian (?) radiolarians *Livarella* sp. The observed thickness is 10 m.

2. Gray and greenish gray, fine-grained sandstones with sinuous bedding and intercalations of light gray siltstones, pink and light red cherts, and pink, micritic limestones with the late Norian radiolarians: *Betraccium* sp. cf. *B. deweveri* Pessagno et Blome, *Livarella longus* Yoshida, *Praemesosaturnalis multidentatus* (Kozur et Mostler), and *Pentactinocarpus* sp. cf. *P. sevaticus* Kozur et Mostler. The thickness is 15 m.

3. Gray and light gray, fine-grained, thin-bedded sandstones, intercalated with reddish gray siltstones. The thickness is 5 m.

EPISKOPI FORMATION

4. Red and brick red, calcitized cherts with numerous thin intercalations of red cherty mudstones, with Callovian–Oxfordian radiolarians: *Gueuxella nudata* (Kocher), *Hsuum brevicostatum* (Ozwoldova), and *Protunuma ochiensis* Matsuoka. The thickness is 20 m.

The correlation of the studied sequences allows us to subdivide the Vlambouros Formation into three main members (Figs. 1 and 2):

1. Middle-grained sandstones with plant detritus and rare beds of micritic limestones, which contain the late Carnian–early Norian conodonts: *Epigondolella abneptis* (Huckriede), and *Neogondolella nodosa* (Hayashi). The presence of middle Norian deposits is not inconceivable. The observed thickness is 10 m.

2. Fine-grained sandstones and siltstones with numerous intercalations of limestones, calcarenites and (sometimes) cherts with the late Norian radiolarians *Betraccium* sp. cf. *B. deweveri* Pessagno et Blome, *Livarella longus* Yoshida and conodonts *Epigondolella bidentata* Mosher. The thickness is less than 20 m.

3. Reddish, fine-grained sandstones and siltstones with rare beds of cherts with the late Norian–Rhaetian (?) radiolarians. The thickness varies from 0 to 20 m.

Member 3 (reddish sandstones and siltstones) was interpreted by previous investigators as the basal layer of the overlying Episkopi Formation (Robertson and Woodcock, 1971; Swarbrick and Robertson, 1980). Lithological and paleontological evidences allow us to include this member into the Vlambouros Formation. This conclusion also concerns the problem of the Marona Formation (monotonous limestones) distinguished in some sections by previous investigators (Robertson and Woodcock, 1979; Swarbrick and Robertson, 1980) as a distinct lithologic unit overlying the Vlambouros Formation in the definition of Robertson and Woodcock (1979), and underlying the red-colored clastic rocks (member 3 of this study). The Marona Formation is a facial equivalent of member 2 of the Vlambouros Formation in our interpretation. The middle part of the Vlambouros Formation is characterized by an abundance of carbonate rocks. Unfortunately, the type section of the Marona Formation is now covered by a landslide. We obtained some information about the structure and position of this carbonate member from a section of the Akamas Peninsula, 2 km south-southwest from the village of the Fasli (Fig. 2), where we distinguished the following units:

VLAMBOUROS FORMATION

1. Gray and yellowish gray, middle- to fine-grained, platy sandstones with plant detritus and thin intercalations of shaly, carbonaceous siltstones. The observed thickness is 10 m.

2. Pinkish gray, thin-bedded siltstones with intercalations of greenish grey siltstones. The thickness is 10 m.

3. Pink and white, platy, micritic limestones with styloliths and thin intercalations of yellowish gray mudstones. Limestones contain the late Norian conodonts *Epigondolella bidentata* Mosher, and *E. abneptis* (Huckriede). The thickness is 2 m.

4. Red, thin-bedded mudstones. The thickness is 2 m.

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5. Red, platy cherts with thin intercalations of red and brick red, cherty mudstones. Cherts contain the *Callovian–Oxfordian* radiolarians: *Gueuxella nudata* (Kocher), *Hsuum brevicostatum* (Ozwoldova), and *Protunuma ochiensis* Matsuoka. The observed thickness is 10 m.

All three members of the Vlambouros Formation are present in this section. Members 2 and 3 are relatively thin and free of coarse-grained sediments (sandstones and siltstones). Because of this, the middle member consists of pure limestones. Their thickness is similar to the total thickness of limestones in the middle member of other sections. It seems more expedient to consider the Marona Formation as a facial equivalent of the middle member of the Vlambouros Formation (Fig. 1).

Previous investigators pointed out that the amount of coarse-grained clastics in the sections of the Vlambouros Formation is constantly decreasing from northwest to southeast (Swarbrick and Robertson, 1980). This is correct only to some extent. The Khapotami River section is relatively depleted in sandstones, and more rich in limestones and cherts. Nevertheless, other sections are almost similar in these characteristics, and there are no grounds to speculate on the decreasing content of coarse-grained clastics from northwest to southeast, because the only notable exception is the Khapotami section. Moreover, the lower part of the Vlambouros Formation in the Dhiarizos River section is represented by the most coarse-grained facies among all of the sections. This probably indicates that the Vlambouros Formation has a more complicated facial structure than was supposed. Its facies variations well fit the model of turbidite fans. Rare and thin intercalations of radiolarian cherts can be interpreted in this case as background sediments of a relatively deep basin.

The middle and upper members of the Vlambouros Formation are characterized by the typical late Triassic radiolarian assemblage, which includes the following taxa (Plates I, II): *Betraccium* sp. cf. *B. deweveri* Pessagno et Blome, *Neopaurinella* sp. cf. *N. sevatica* Kozur et Mostler, *Livarella longus* Yoshida, *L.* sp. cf. *L. densiporata* Kozur et Mostler, *L. gifuensis* Yoshida, *L.* sp., *Pentactinocarpus* sp. cf. *P. sevaticus* Kozur et Mostler, *P. magnus* (Kozur et Mostler), *Praemesosaturnalis* sp. aff. *P. gracilis* (Kozur et Mostler), *P. multidentatus* (Kozur et Mostler), *P.* sp. aff. *P. multidentatus* (Kozur et Mostler), *P. quadriradiatus* (Kozur et Mostler), *P. zapfei* (Kozur et Mostler), *P.* sp. I, *P.* sp. II, *P.* sp. III, *Tetraspinocyrtis* (?) sp., and multirayed *Livarella*-like forms. It is a taxonomically typical assemblage of the late Norian, which can be correlated with coeval assemblages from the Alps (Kozur and Mostler, 1972, 1979, 1981), eastern Russia (Bragin, 1991), Japan (Yao, 1982; Yoshida, 1986), and western regions of North America (Blome, 1984). The assemblage is typical of the upper Norian–Rhaetian radiolarian zone of *Canoptum triassicum* (Bragin, 1991), or the upper Norian zone of *Betraccium deweveri* (Blome, 1984; Carter, 1993). The low taxonomic diversity of this assemblage should be mentioned. The representatives of the Pantanelliinae subfamily are very rare (with the exception of the genus *Betraccium*). Cyrtoid nassellarians are also rare, and the genus *Canoptum*, which is typical of this time interval, is absent. There is a certain quantitative predominance of the genera *Livarella* and *Praemesosaturnalis*. It may be considered as evidence of the neritic character of the studied assemblage. Sometimes, when taxonomic diversity of the assemblage was very low, the age was estimated in a more wide range (as the late Norian–Rhaetian). Unfortunately, the obtained data cannot prove the presence of Rhaetian deposits in the Vlambouros Formation.

Exceptionally interesting and unusual relations between the Vlambouros and overlying Episkopi formations suggest a very large hiatus, which includes a stratigraphic interval approximately from the Hettangian to middle Callovian. This hiatus can be traced in all studied sections and considered as regional. Two versions can be proposed for the interpretation of this phenomenon: (1) the gap in sedimentation during the Early and Middle Jurassic resulted from abissal conditions similar to those in recent deposition areas of red oceanic clays, with a very low sedimentation rate; (2) the erosion of Lower and Middle Jurassic deposits. We have to estimate the probability of these versions by analysis of the lithologic evolution of the Vlambouros Formation on the basis of new stratigraphic data.

The lower member of the Vlambouros Formation is represented by an alternation of thick-bedded, sandy turbidites. Their separate rhytms can have a thickness of 1 m. In this case, the A–C intervals of the Bouma turbidity cycle (Bouma, 1962) are distinct. More fine-grained sandstones with typical turbidity structures showing elements B–C, B–C–D and more rarely B–C–D–E, appear in the middle part of the lower member. The micritic limestones, which can be interpreted as the horizon E of a turbidite cycle or the background basin sediments (Walker, 1967), appear at the same level. The upper part of the lower member is represented by the alternation of middle- to fine-grained sandstones with calcilutites and calcarenites. The elements C–D–E are most common here, and the elements B–C–D and A–B–C are rare. Generally, this part of the section shows gradual decreasing of the thickness of separate rhythms (sandstones beds), gradual replacing of coarse- and middle-grained sandstones by fine-grained clastic rocks (horizons A and B of Bouma cycle become more rare), and the appearance of calcarenites in the upper part of the member. It should be noted, that some sediments of the lowermost part of the Vlambouros Formation, which are common at the Akamas Peninsula, were interpreted by several authors as very shallow, deltaic deposits, the products of a very quick sedimentation (Robertson and Woodcock, 1981).

Petrographic data show that sandstones of this member can be discriminated in quartz subarkoses, litharenites and calcarenites. Subarkoses are composed of angular and semi-rounded quartz grains with subordinate grains of plagioclase, K-feldspar, perthite, hornblende, epidote, mica, and rare ore and heavy minerals. Some quartz grains have an Fe-oxide coating that suggests a long history of transportation or multistage redeposition. The cement consists largely of clay minerals (mostly hydromicas) building up to 20–30% of the rock volume. The features allow us to refer these sandstones to the subarkose-vacca type (Middleton, 1970; Middleton and Hampton, 1973). Apart from quartz, feldspar and micas, litharenites contain angular fragments of carbonate rocks, acid volcanics, quartzite and micaceous schists; their intervals B and C bear fragments of coalified wood. Carbonate fragments often appear in various elements of the turbidity cycles, giving rise to calcarenites and lithocalcarenites. Oncolitic sandstones should also be noted among these petrographic types. They form thick beds (up to 1 m) composed of several layers, with a thickness range of 10–15 cm. The oncolite cores are represented by quartz grains or the aforementioned rock fragments. This structure of oncolites suggests that they formed in the shallow-water area, and were redeposited by turbidity or autokinethic flows (Walker, 1967). The small percent of oncolites is detectable in all petrographic types of sandstones from the lower member of the Vlambouros Formation.

The middle member of the Vlambouros Formation is distinguished by an abundance of micritic limestones, calcarenites and cherts, which form numerous intercalations among sandstones and siltstones. Sandstones of this member are petrographically similar to this rock type in the lower member, and are predominantly of turbiditic origin. In addition to turbidity sandstones, this sequence contains fine-grained varieties with very fine sinuous to horizontal bedding and very sharp contacts between separate layers displaying a very high degree of granulometric sorting. All of these features, as well as the basal structures, can be interpreted as the result of the redeposition of these sediments by bottom currents (Curray and Moore, 1974; Hampton, 1972; Hein, 1982; Hollister *et al.*, 1974; Hubert, 1964). Such sandstones are determined to be contourites. We have to mention that the petrographic composition of contourites and turbidites is very similar. It should also be noted that oncolitic sandstones are very rare at this level, and are completely absent among contouritic sandstones. The upper parts of turbidity cycles often contain micritic limestones, representing the interval E. These limestones are usually pink in color, and contain numerous remains of radiolarians and scattered shells of halobians. These features, as well as the position of the carbonate intercalations relative to turbidity cycles, indicate that micritic limestones are the products of background sedimentation, which was disturbed by the parallel formation of turbidites and contourites.

The upper member of the Vlambouros Formation has an essentially different lithology. This member is characterized by the thin interbedding of fine-grained sandstones, siltstones, mudstones, calcarenites and calcilutites. Micritic limestones are rare. Here, there are two types of litharenite beds. The first one is represented by thin beds (2–5 cm) with sinuous- to cross-bedding and good sorting of the material, while the graded bedding is missing. Abrupt, contrasting bedding with sharp sandstone-siltstone boundaries as well as flow grooves are typical. The second type is represented by thicker (5–10 cm) beds, which may show C–D and B–C–D Bouma intervals. In addition to indistinct graded bedding, there are pendulum, cross, and criss-cross bedding in the last rocks. Horizons of sandstons with very thin parallel bedding sometimes occur. Laminations of a very small scale (1–2 mm) can be traced

in such beds at a distance of several dozens of centimeters. The petrographic composition of all these sandstones is identical to that in the underlying clastic rocks. It should be noted that calcarenites are more widespread in the upper member. These rocks are predominantly fine- and medium-grained. They contain clastic fractions (0–30%) of quartz, feldspars, acid volcanics, quartz-mica schists, quartzites, microquartzites and siliceous rocks. Carbonate material is represented by micritic grains, pellets, radiaxial and tangential oolites, recrystallized foraminifers, shell detritus and sponge spicules. This material has an undoubtedly shallow-water origin. Intraclasts are represented by fragments of organoclastic limestones, marls and micritic limestones.

Calcilutites usually have fine laminar-flow structures due to the changing grain size from micritic to calcarenitic. Graded bedding is typical of thin layers, which sometimes contain coarse-grained carbonate or terrigenous clastic material and shell detritus (halobians?). Terrigenous material is represented by rare grains of quartz and plagioclase, and by more frequent mica flakes oriented parallel to bedding planes. Carbonate material is represented mostly by calcite grains. Carbonate micritic grains are rare, scattered and contain radiolarian remains, replaced by calcite. Calcilutites exhibit transitions to the micritic limestones, which contain fine-grained carbonate material composed of shell detritus and scattered calcitized radiolarian fragments. Preserved halobian shells are oriented along bedding planes. Some types of micritic limestones are bioturbated. They show clotty structures and calcitized feeding burrows. The upper parts of calcilutite beds usually exhibit silicification and gradual transitions to the radiolarian cherts. They can be interpreted as background sediments of a basin with normal salinity and aeration.

The petrography of clastic sediments of the Vlambouros Formation suggests that erosion affected the supracrustal metamorphic complex and was accompanied by the redeposition of shallow-water (shelf) carbonate sediments. Erosion of the metamorphic complex decreased with time, while the redeposition of carbonate sediments gained more importance. Background "pelagic" sedimentation also took place, forming radiolaria-bearing micrite and chert beds. This sedimentation was of secondary importance. All these data allow us to conclude that the Vlambouros Formation was deposited in the marginal part of a basin, relatively close to the continental block surrounded by carbonate platforms. These conclusions are concordant with an idea suggesting that the Vlambouros Formation and its equivalents in other regions (northwestern Syria) are sediments of the southern passive margin of the Tethys (Delaune-Mayere *et al.*, 1977). Nevertheless, the exact position of the Vlambouros deposits in the southern areas of the Tethys is still questionable (Kaz'min *et al.*, 1987; Robertson, 1990; Robertson and Woodcock, 1981).

On the other hand, the development of sedimentation corresponding to the studied deposits of Cyprus consist in the gradual transition from proximal to more distal turbidites, and then to contourites. This results in the erosion and redeposition of turbiditic sediments by currents. Such a formational change can be caused by rifting, which was the main reason for the general widening and deepening of the Mesotethys and its continental margins. Some more general reasons may also be important. The deposition of the Vlambouros Formation is coeval to the large eustatic transgressive cycle (Reding, 1990). This is concordant with the decrease of clastic material upsection, and the more wide development of the carbonate shelf areas, which were the provenance for the calciturbidite sediments of the Vlambouros Formation. The general transgression and widening of the basin probably changed the current patterns and caused the erosion of some sediments. This can be supported by vertical changes in the section lithology and the inconsistency of the upper member of the Vlambouros Formation, which is completely absent in some localities. It seems to be very reasonable that the stratigraphic hiatus between the Vlambouros and Episkopi formations is a result of bottom current activity. At the same time, it should be noted that a hiatus, when caused only by currents, cannot span the entire Lower and Middle Jurassic intervals everywhere in the wide area. The study of the Lower and Middle Jurassic deposits of the Mamonia Complex or its equivalents in the Eastern Mediterranean is of primary importance for solving the problem of the hiatus and for a better understanding of the Mesotethys evolution.

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The Upper Eocene of the Fergana Depression and Adjacent Regions

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Abstract—Radiolarians and nannofossils have been studied in the two most complete upper Eocene sections of the Fergana depression, the classical region of the shallow-water Paleogene deposits in the southern part of the former Soviet Union (FSU). The sections display three periods of the Fergana basin deepening and its more open connections with a main aquatic region: the second half of the Rishtan time, and the Isfara and Sumsar periods. Three zones of the standard Eocene zonation are distinguished for the first time in the Fergana depression: the radiolarian *Thyrsocyrtis bromia* Zone in the Isfara deposits, and the nannoplankton zones NP19–NP20 in the Sumsar deposits. Two detected transgressive–regressive cycles of sedimentation comprise the late Rishtan–Khanabad and Sumsar intervals. The transgressive part of the former is correlative to the early Priabonian eustatic transgression. The second cycle comprises only transgressive deposits, and is correlated with the late Priabonian (Lattorfian) eustatic transgression.

Key words: *Paleogene, Central Asia, stratigraphy, micropaleontology.*

The Fergana depression is a well-known region of the FSU, where the shallow-water Paleogene deposits display a relatively complete stratigraphic succession, are well exposed, and contain diverse macrofossils (mostly molluscs). Sediments here accumulated under conditions of a past marine bay, which had restricted, yet uninterrupted connections with the main body of water, the Turan Sea. This sea was a part of a vast system of interrelated epicontinental basins of the Crimea–Caucasus, Central Asia, and adjacent regions. The southern (respectively more warm) regions of the system were the areas of predominantly carbonate sedimentation with carbonate components produced, as a rule, by nannoplankton and foraminifers. The periods when biota in these regions were rich in siliceous microorganisms (radiolarians, diatoms, and sponge spicules) were extremely rare. Macrofauna provided an insignificant contribution to the carbonate fraction of sediments.

The distribution of plankton in the southern basins was controlled by their depth. Among them, the Fergana basin was of the shallow type, countered the prosperity of plankton (both the calcareous and siliceous), and thus favored the macrofauna (molluscs, first of all *Ostrea* forms) to be the main contributor of carbonate material to the sediments. However, sometimes, when the water exchange between the bay and the main body of water was more or less free, the mineral remains of plankton organisms became the notable components of sedimentation.

Initial studies on the Paleogene sections in the Fergana depression date back to early in the last century, and one of their results was Vyalov's scheme (1934) of the so-called Central Asian stages of the Paleogene based on the distribution of the *Ostrea* assemblages. For a long time, this scheme was a classic example considered as evidence of the regional stage succession, which is independent on the general chronostratigraphic scale. Attempts to apply the Central Asian scheme were widely practiced (up to eastern Turkmenistan) before the early 1960s, when its advocates finally came to a conclusion which was originally obvious: when based on a particular feature, the stratigraphic subdivisions are valid only in that area, where this feature is observable. In our case, this means that subdivisions, whose boundaries are substantiated by using molluscan assemblages, are recognizable only inside the distribution area of these assemblages (precisely the assemblages, but not their constituent forms), i.e., in the Fergana and Tajik depressions along with regions closely adjacent to them.

The formerly known Paleogene assemblages of microfauna from the Fergana depression (basically foraminifers and radiolarians) are poor and inapplicable in recognizing stage subdivisions in the relatively deep water facies, and for correlating these units with the general stratigraphic zonation of the Paleogene. As a result, the stage ages changed many times since the moment when they were first recognized.

Nevertheless, the Paleogene in the Fergana area is a unique object to study (first of all, with respect to its

malakofauna). The succession of its subdivisions, which was distinguished more than 50 years ago, is still the basis for stratigraphic correlation not only here, but also (when modified) in the Tajik depression and other regions nearby—i.e., in the vast territory of Central Asia, although at present, these subdivisions are ranked as horizons and formations, not as stages.

In this work, we focused our attention first, on the revision of the radiolarian assemblages from the upper Eocene deposits of the Fergana depression (they appeared to be more diverse and informative than was earlier believed), and on new finds of nannoplankton remains from the same stratigraphic interval. This enabled us to distinguish for the first time the general zonation units in the upper Eocene sections of the depression. Secondly, we analyzed the upper Eocene sedimentary succession of the Fergana depression with regard to its transgressive–regressive cycles of regional, interregional, and global (eustatic) rank in order to demonstrate that the succession itself and its intervals bearing plankton remains are not the results of local factors, but the consequences of events of a higher rank.

The paper reviews data of the long-term study on the Paleogene of the Fergana depression by the Chair of Paleontology at Tashkent University under the supervision of Rasulov. Muzylev assisted in studying the Isfara and Shor-Su sections of the depression that provided data important for this work. Radiolarians and nannofossils from these sections were identified by Khokhlova, and Muzylev, in collaboration with Kushakov, respectively.

The stratigraphic interval of our interest (along with underlying and overlying beds) includes the traditional Rishtan, Isfara, Sumsar, and Massaget horizons. Their ages were corrected many times with parallel changes in the local stratigraphic range of the upper Eocene. During the last decade, it was concluded that the Eocene lower boundary corresponds to the level dividing the lower and upper Rishtan subhorizons (Salibaev, 1983; and others). Even earlier, the molluscan fauna from the Sumsar Horizon was proven to be correlative to its representatives from the Lattorfian Stage of northern Germany, and since the latter stage is universally accepted as late Eocene in age, it is common practice to believe that the Sumsar deposits are terminating the upper Eocene succession in the Fergana depression.

Below, we present brief characteristics and our comments on the above stratigraphic units in the Isfara and Shor-Su sections (Fig. 1).

The Rishtan Horizon. All over the depression, this horizon is divisible into two units: the lower subhorizon comprising deposits of shallow-water origin, and the upper one composed of deep-water sediments. In the Isfara and Shor-Su sections, the lower Rishtan Subhorizon is distinct as a unit of light-colored normal and silty sands, with coquina beds up to 10 m in thickness. The upper subhorizon consists of brownish gray slightly calcareous clay enclosing interlayers of oyster shells up to 30 m in thickness. There is a great difference between the mollusks from lower and upper units, and Salibaev (1983) suggested that the former Rishtan Stage be discarded in favor of distinguishing the Ichkibulak (lower) and Tair-Su (upper) subhorizons as two independent units instead.

The Isfara Horizon. This horizon of 25–30 m in thickness is composed of light to brownish gray carbonate-free siliceous clays and radiolarites bearing abundant fish remains.

The Khanabad Horizon. Sediments of this horizon show the gradual transition to underlying deposits. They are represented by green carbonate-free silty clays, whose total thickness is approximately 30 m. On the basis of the gradual transition between the Isfara and Khanabad horizons, which is combined with their obvious similarity in paleontological and palynological characteristics, Davidzon (1969) suggested to consider them as the indivisible Kushan Horizon; this was proved to be reasonable by the detailed study on molluscan fauna from both levels (Salibaev, 1983; and others).

The Sumsar Horizon. This unit has a well traceable boundary bed of crimson calcareous clays with a basal interlayer of phosphatic nodules and casts of molluscan shells. Some Fergana sections display pre-Sumsar erosion marks, suggesting that the Sumsar sediments transgressively overlap the older deposits, including those of the Turkestan level (Salibaev, 1982).

The Massaget Horizon. This horizon of red silty clays with silt interlayers and a basal bed of green sand unconformably rests on a distinct erosion surface.

Prior to our discussion on fossil plankton assemblages from the upper Eocene deposits of the depression, we would first like to outline the current situation with radiolarian and nannoplankton zonation for this stratigraphic interval.

In the standard zonation of tropical radiolarians, the upper Eocene corresponds to the *Thyrsocyrtis bromia* Zone (Riedel and Sanfilippo, 1978). The more detailed zonation scale (Palmer, 1988) suggests that this interval includes the *Carpocanistrum azyx* and *Calocyclas bandyca* zones, however, these two units are not traceable everywhere within the tropical belt. In the radiolarian scale for the temperate boreal latitudes, the upper Eocene is referred to the *Theocyrtis andriashevi* Zone (Kozlova, 1990).

On the basis of nannofossils, the upper Eocene is divisible into three units: NP18 or the *Chiasmolithus oamaruensis* Zone, which also comprises the uppermost part of the middle Eocene; NP19 or the *Isthmolithus recurvus* Zone; and NP20 or the *Sphenolithus pseudoradians* Zone. In southern regions of the FSU, it is difficult to identify the boundary between the two latter zones (specimens of *Sph. pseudoradians* are either extremely rare, or lacking here), and it is a common

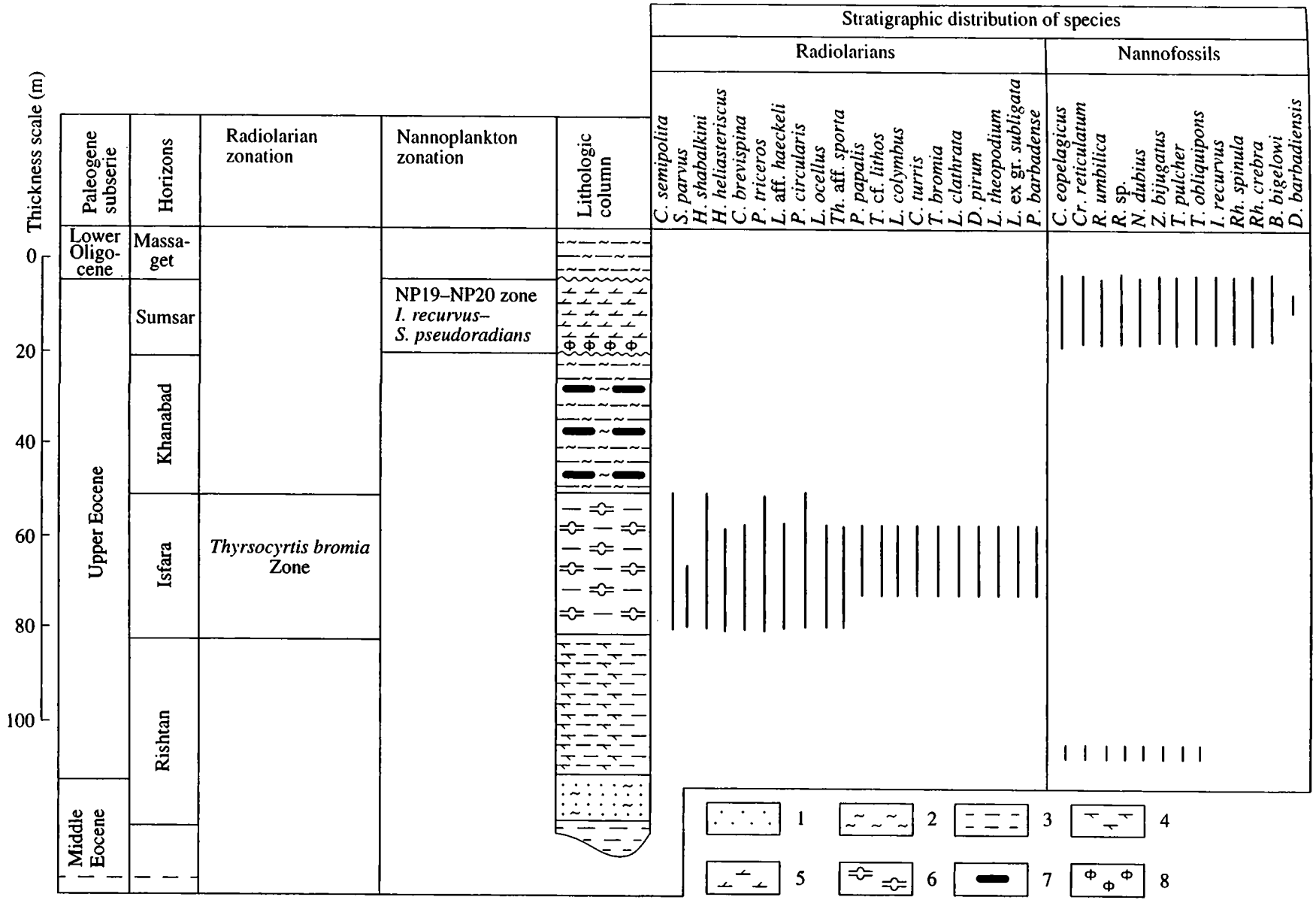

Fig. 1. Composite upper Eocene section for the Fergana depression with stratigraphic zonation of radiolarian and nannofossil species: (1) sand and sandy fractions; (2) siltstone and silty fractions; (3) clay and clay fractions; (4) low-calcareous clay; (5) calcareous clay; (6) siliceous clay; (7) concretions; (8) phosphatic nodules.

practice here to distinguish the undivided NP19–NP20 zones interval.

The problem of special interest concerns the position of the Eocene–Oligocene boundary in the nannoplankton zonation scale and in sedimentary sequences occurring in the southern part of the FSU. This problem is important both in the general aspect, and in the correct stratigraphic subdivision of the Fergana sections.

The Eocene–Oligocene boundary standard is thought to be the boundary between the foraminifer zones P17 and P18. Earlier, this boundary was considered as correlative to the boundary between the nannoplankton zones NP20 (*Sphenolithus pseudoradians*) and NP21 (*Coccolithus subdistichus*), which is marked by the disappearance level of *Discoaster barbadiensis* and *D. saipanensis*. The top of the NP20 zone is the most unreliable datum level in the Cenozoic nannoplankton scale, because both index species are of the thermophilic type, and their geographic and stratigraphic distribution is under strong temperature control. Specimens of *D. barbadiensis* and *D. saipanensis* are usually rare in the upper part of the NP20 zone, and may be totally lacking in deposits of temperate latitudes. This seems to be the reason why in the most recent works, the upper boundary of the NP20 zone is placed at a level lower than the top of the foraminifer zone P17. The special study on foraminifers and nannofossils from the continuous sections of the northern Caucasus has shown that, despite the extremely rare occurrence of *D. barbadiensis* and *D. saipanensis* in the terminal upper Eocene beds, their last occurrence level is still identical to the Eocene–Oligocene boundary based on the distribution of foraminifers, and coincident with the boundary between the Belaya Glina Formation and Maikop Group (Krasheninnikov and Muzylev, 1975; Muzylev and Tabachnikova, 1987).

Abundant radiolarians are thus far known from the Isfara deposits, while their rare and poorly preserved specimens were also reported from lower beds of the Khanabad Horizon. Lipman (1953) was the first to study the Isfara radiolarian assemblage and to distinguish some new species (for instance, *Heterosestrum tschuenkoi* Lipm. and *H. shabalkini* Lipm.) which appeared to be forms of distinct stratigraphic significance. The Isfara assemblage was also studied by Tschediya and Zainutdinov (1973). Among the new forms described by these researchers, the *Polydiscus compositum* Zajn. species and varieties of *Cenodiscus hexaporicus* Tsched. et Zajn. were referred to the guide fossils.

Progressive maceration methods and numerous additional data on the worldwide distribution of the late Eocene radiolarians afford an opportunity to revise their assemblage from the Isfara Horizon against this new background of our knowledge.

Radiolarians are typical of the total horizon interval in both the Isfara and Shor-Su sections, although their abundance, diversity, and degree of preservation are inconstant. Lower beds of the interval yielded only poor assemblages of moderate and low preservation degree. These assemblages are most rich in discoids (genera *Ommatodiscus*, *Heterosestrum*, *Heliodiscus*, and *Stylodyctya*) and sphaeroids. Nasselarians are represented here by the genera *Lithomelissa* and *Theocorys*. The forms, which were identified at the species level, include *Cyrtophormis brevispina* Vinassa, *Calocyclas semipolita* (Clark et Campbell), *Lithomelissa haeckeli* Butschlii, *Sethoconus parvus* Gorbunov, *Petalospyris triceros* (Ehr.), *Heterosestrum shabalkini* Lipm., and *Plectodiscus circularis* (Clark et Campbell). In general, these are temperate-boreal and cosmopolitan species characteristic of a wide stratigraphic range (middle to lowermost upper Eocene). The only exception is *Heterosestrum shabalkini*, first described by Lipman precisely from the Isfara beds of the Fergana depression, and next reported from a lower part of the upper Eocene sediments of the Voronezh anticlise (Tochilina, 1972), from the upper Eocene deposits of the Russian platform (Kozlova, 1990), and from presumably upper Eocene strata of the Norway–Greenland region (Kozlova and Petrushevskaya, 1979).

Deposits of the middle part of the horizon yielded the rich and diverse (especially in the Isfara section) assemblage of ideally preserved radiolarians (more than 20 species). In addition to the aforementioned cosmopolitan forms and species characteristic of epicontinental basins, this assemblage includes the following oceanic radiolarians: *Lamptonium colymbus* Foreman, *Podocyrtis papalis* Ehr., *Thyrsocyrtis bromia* Ehr., *Pterocorythidium barbadense* Riedel et Sanfilippo, *Dictyoprora pirum* (Ehr.), *Calocyclas turris* Ehr., *Lirospyris clathrata* (Hck.) and other representatives. The listed species definitely indicate the late Eocene age of host deposits, and more precisely, their coherence with the *Thyrsocyrtis bromia* Zone of the standard radiolarian scale.

In upper beds, the radiolarian assemblage is again poor, and its representatives are partially dissolved.

It is also remarkable that the Isfara Horizon in both studied sections also contains diatoms, locally very abundant. The diatoms show a high taxonomic diversity despite the low and moderate preservation degree. Their assemblage includes the late Eocene index species, though these forms need to be studied in more detail (E.P. Radionova, personal communication).

Nannofossils are confined to two stratigraphic levels of the Isfara and Shor-Su sections. The basal upper Rishtan beds of the Isfara section yielded rare specimens of *Coccolithus eopelagicus* (Braml. et Riedel), *Cribrocentrum reticulatum* (Gartner), *Reticulofenestra umbilica* (Levin), *Neococcolithes dubius* (Defl.), *Transversopontis pulcher* (Defl.), *Tr. oblicuipons* (Defl.), and *Zighrablithus bijugatus* (Defl.). Unfortunately, this assemblage is lacking the zonal species. Its *Cr. reticulatum* form suggests that the deposits are not older than the second half of the middle Eocene, and

this neither clarifies the available viewpoint on the stratigraphic position of the Rishtan Horizon, nor discredits it.

A more diverse nannofossil assemblage was found in the Sumsar Horizon of both sections. In addition to above species, it also includes *Isthmolithus recurvus* Defl., *Rhabdosphaera spinula* Levin, and more sporadic specimens of *Rh. crebra* (Defl.), *Braarudosphaera bigelowi* (Gran et Braar.), and *Discoaster barbadiensis* (the latter as a single specimen from the Isfara section). This assemblage is typical of the undivided NP19–NP20 (*Isthmolithus recurvus–Sphenolithus pseudoradians*) zone of the upper part of the upper Eocene.

Thus, the studied interval yielded three assemblages of fossil plankton forms: the nannofossil assemblage of unclear zonal affinity from the base of the upper Rishtan Horizon, the radiolarian assemblage of the *Thyrsocyrtis bromia* Zone of the upper Eocene from the Isfara interval, and the nannofossil assemblage of the undivided NP19–NP20 (*Isthmolithus recurvus–Sphenolithus pseudoradians*) zone from the Sumsar Horizon. Since the latter horizon is the uppermost in the respective succession, it seems reasonable to suggest that this unit generally corresponds to an upper part of the NP19–NP20 interval, i.e., to the NP20 zone.

It is intriguing to compare these stratigraphic conclusions with data on nannofossils from coeval deposits in the nearby Tajik depression. In the Paleogene geological history, both the Fergana and Tajik basins were closely related, although the latter was notably deeper, and its connections with the main body of water were less restricted. As a result, the Tajik basin was more rich in plankton than the Fergana depression. The upper Eocene of this basin, together with boundary beds below and above, is subdivided into the Ichkibulak (=lower Rishtan in the Fergana depression), Tair-Su (=upper Rishtan), Kushan (=Isfara + Khanabad), Sumsar, and Gissarak (=Massaget) horizons. In terms of the lithologic features, these deposits of deeper sedimentation settings are similar to their counterparts in the Fergana basin.

The Ichkibulak Horizon is barren of nannofossils. Their poor assemblage from the Tair-Su Horizon includes only cosmopolitan species. That part of the Kushan deposits, which corresponds to the Isfara Horizon, again is free of these fossils. A preliminary study on nannofossils from the upper Kushan beds has shown that they are of low diversity and do not include zonal forms (Muzylev and Salibaev, 1988), although later Muzylev (unpublished data) identified in this assemblage rare specimens of *Isthmolithus recurvus*, the index species of the NP19–NP20 zone. Similar representatives were also distinguished in the Sumsar deposits, where nannofossils are more diverse and significantly more abundant.

In turning now to the analysis of the upper Eocene sedimentary sequence from the viewpoint of transgressive–regressive cycles of sedimentation, we should pay particular attention to episodes and periods of the siliceous and calcareous plankton invasions into the shallow Fergana bay. Since biota of the bay was generally lacking the plankton organisms, their ingress should be considered as indicating a phase of deepening and closer links with the main body of water.

The above data point to three periods of plankton ingress during the early Rishtan (nannofossils), Isfara (radiolarians), and Sumsar (nannofossils) times.

The idea of subdividing the Paleogene sections of the Fergana and Tajik depressions into regional transgressive–regressive cycles of sedimentation goes back to the 1950s, when Simakov (1953) recognized three of them, and identified the latest one as corresponding to the upper Rishtan–Sumsar interval. The incorrect concept of considering the Sumsar Horizon as exemplifying the regressive deposits is widespread, and it is now difficult to guess which of original facts was misleading. It is likely that it was the crimson color (traditionally attributed to terrestrial sediments) of the Sumsar deposits, because all other facts suggest, in contrast, the transgressive environments of their accumulation. Such facts include the basal eroded surface transgressive overlapping of older deposits, basal phosphatic nodules, and abundance of plankton remains.

Therefore, the late Eocene of the Fergana depression should comprise not a single, but two transgressive–regressive cycles corresponding to the late Rishtan–Khanabad and Sumsar periods. The beginning of the first cycle is evident from the transition of the lower Rishtan sands into the upper Rishtan poorly calcareous clays, whose basal beds contain scarce nannofossils (although separate oyster-shell accumulations in these clays still suggest the shallow depths of the late Rishtan basin in general). The peak of transgression was in the Isfara time (fine clays with abundant radiolarians, many of which are the "oceanic" forms). A transition from this phase to the new regression is detectable at the beginning of the Khanabad interval (the basal Khanabad beds still contain scarce radiolarians). The regression peak is confined to the pre-Sumsar erosion. The next cycle is reduced and represented only by the transgressive Sumsar semicycle (calcareous clays with abundant nannofossils); its regressive part corresponds to the pre-Massaget hiatus in sedimentation. It is not inconceivable that additional studies will bring forth data critical for subdividing the lower cycle into two subcycles corresponding to the late Rishtan and Isfara–Khanabad intervals (see below).

Our next objective is to reveal transgressive–regressive cycles in other Paleogene basins of the Crimea–Caucasus, Central Asia, and adjacent regions (Fig. 2). For the Volgograd region near the Volga River, this problem is of particular interest (Marzuk *et al.*, 1992), because Paleogene deposits of this area contain well-studied and stratified nannofossil and foraminifer assemblages. In addition, paleobasins in this area were

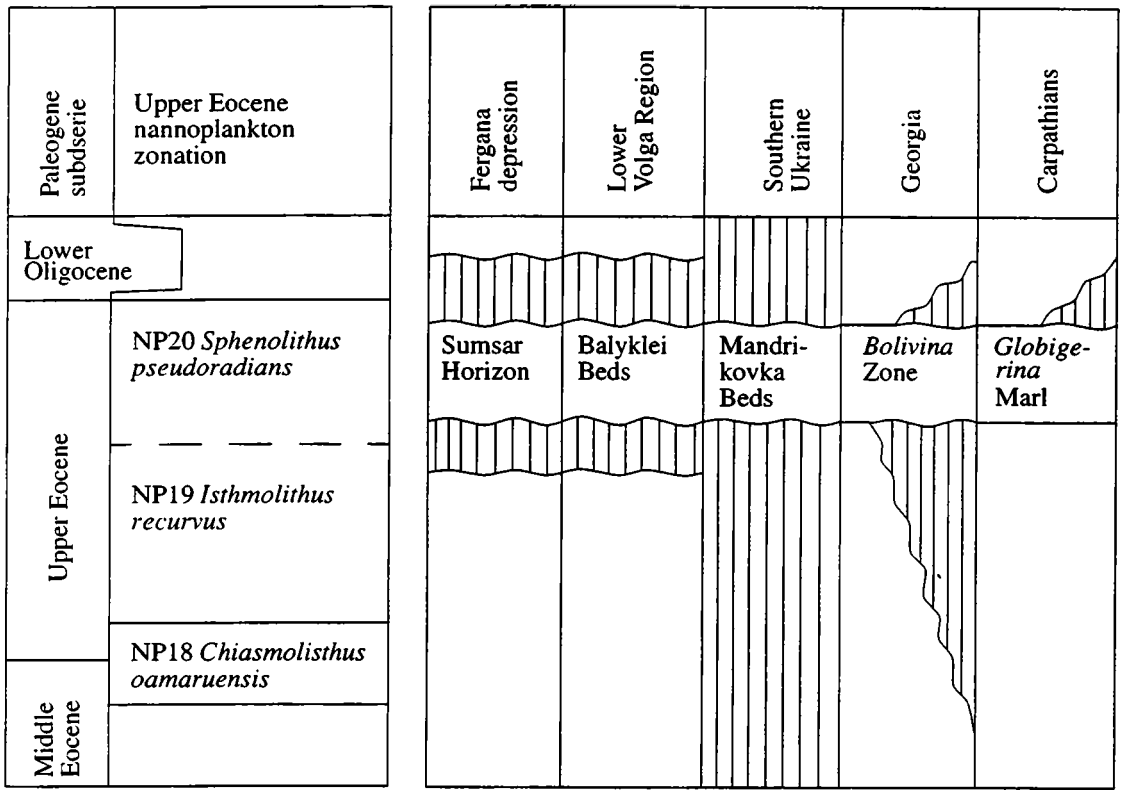

Fig. 2. Level of the Lattorfian (latest Eocene) transgression in southern region of the former Soviet Union.

not very deep, and transgressive–regressive phases of their evolution are clearly recorded in sedimentary sections. In particular, the middle–upper Eocene sequence here comprises mostly transgressive sedimentary facies, whereas its distinct lithologic and stratigraphic hiatuses correspond to regressive phases of the basins' evolution. The upper Eocene portion of the sequence includes two transgressive units of nanno-foraminiferal marls and calcareous clays, both bounded by hiatuses. The lower unit corresponds to the undivided NP19–NP20 zone, whereas the upper one is equivalent to the NP20 zone, as is indicated by the index species discovered there. The last unit (up to several meters in thickness) is known as the Balyklei Beds; with a significant hiatus, they rest on the middle Eocene deposits, and are overlain, again with a hiatus, by the lower Oligocene strata.

The latest Eocene transgression is also clearly recorded in other southern regions of the FSU, for instance, in the well-known Mandrikovka Beds near Dnepropetrovsk. These organogenic-detrital carbonate sediments (few meters in thickness) overlap the Early Precambrian granites, and are overlain by the Quaternary loam. The beds contain nannofossils, whose assemblage is typical of the undivided NP19–NP20 zone.

In many sections of Georgia (Kacharava and Khuchua, 1991), transgressive overlapping is characteristic of deposits bearing foraminifers of the *Bolivina* Zone (approximately corresponding to the nannoplankton zone NP20). Distinct traces of the latest Eocene transgression have been observed in the Carpathians, where the Bystritsa Formation of the upper Eocene is composed of terrigenous deposits. The overlying Menilite Formation of the Oligocene age (the unit correlative to the Maikop Group of more southern regions) consists of siliciclastic deposits variably enriched in organic matter. The *Globigerina* Marl below the Menilite Formation base is traceable nearly everywhere in the Carpathians as the unit corresponding to the latest Eocene transgression. It is a few meters thick and very rich in foraminifers and nannofossils. The boundary between zones NP20 and NP21 is thought to be inside this *Globigerina* Marl (Van Couvering *et al.*, 1981), yet we already argued above that it should be displaced higher, up to the base of the Menilite Formation.

Thus, traces of the late Eocene transgression, which have been observed in the Fergana and Tajik depressions, southern Ukraine, lower Volga regions, Georgia and the Carpathians, seem to be synchronous everywhere according to the distribution of nannofossils.

It is intriguing to compare the distinguished transgressive–regressive cycles with those of the eustatic origin. As we know, there were three eustatic transgressions and related regressions during the late Eocene time (Haq *et al.*, 1987; Keller *et al.*, 1987). The last of

them, like similar events in the regions under consideration, is correlative with zone NP20.

The situation with the two other cycles is not so clear. The middle eustatic transgression corresponds to the middle of zone NP19, and the earlier one was in the late time interval of preceding zone NP18. However, available data on the stratigraphic ranges of these zones are controversial: according to Haq and his colleagues (see eustatic curve in their work), zone NP18 is approximately equal to zone NP20, while Keller *et al.* showed it as exceeding the summary range of both the NP19 and NP20 zones. In our opinion, which is based on geological records in the Crimea and northern Caucasus, where Priabonian deposits are represented by a thick sequence of nanno-foraminiferal marls accumulated under a constant sedimentation rate, zone NP18 is 4–5 times lesser than the NP19–NP20 range. In this case, the early and middle Priabonian transgressions would be both inside the range of the CP15b zone. On the other hand, only one transgression of the NP19–NP20 interval is detectable in sections of the Volga region. Consequently, it is not inconceivable that in southern regions of the FSU, the early and middle Priabonian transgressions were manifested as a single event.

Turning back to the upper Eocene deposits of the Fergana depression, we should remind that they display the complete transgressive–regressive cycle of the late Rishtan–Khanabad time, and the transgressive semicycle of the Sumsar time. The above discussion suggests that these events exceed the regional rank. There is a high probability that the late Rishtan–Isfara semicycle corresponds to the early Priabonian transgression in the Volga region. If this is correct, then basal beds of the upper Eocene are most probably lacking in the Fergana depression. As for the Sumsar deposits, they were surely accumulated during the latest eustatic transgression of the late Eocene time.

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The Middle–Late Pleistocene Stratigraphy and Paleogeography in Rybinsk–Yaroslavl Region near the Volga River

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Abstract—New results of a comprehensive investigation are used to substantiate the stratigraphic position of marker beds in the Rybinsk reference section, and to analyze periodic changes in paleogeography during the middle–late Pleistocene. Different glacial and interglacial deposits widespread in reference sections of the Yaroslavl region adjacent to the Volga River are discriminated on the basis of mineralogical, petrographical, and palynological criteria. Correlative lithologic features of the Dnieper, Moscow, and Kalinin moraines, which differ in occurrence mode, inner structure, and composition, are depicted to solve problems concerning these controversially interpreted deposits. Their stratigraphic relations to the Mikulino and early Valdai limnic deposits are outlined. Relevant data of spore–pollen analysis are used to reconstruct the landscapes and climatic conditions that existed during the entire stratigraphic interval of the Mikulino interglaciation. The documented traces of two early Valdai interstades are preceding the latest glaciation in the region, and suggest the late Pleistocene (Kalinin) age of the upper moraine.

Key words: *stratigraphy, paleogeographic reconstructions, paleoclimatic rhythms, reference sections, morains, lithology, palynology, Mikulino interglaciation.*

The Rybinsk–Yaroslavl region adjacent to the Volga River is the key area for solving the problems of the middle–late Pleistocene stratigraphy and paleoclimatic rhythms. The area comprises several well-known localities of the Rybinsk (early to late Pleistocene), Chermenino (Mikulino interglaciation–early Valdai interval), Yakovka (Mikulino and Kalinin horizons), and Dolgopolka (middle–late Pleistocene) reference sections (Arslanov *et al.*, 1974; Valueva *et al.*, 1969; Novskii, 1975; *Razrezy ...*, 1977; Spiridonova *et al.*, 1976; Sudakova, 1990). Regional stratigraphic problems, which have caught the attention of many researchers, can be briefly indicated as follows: the controversial stratification and interpretation of glacial deposits, including discussion on the Kalinin moraine; ambiguous identification of the Likhvin and comparable Bezhitsk deposits; and underestimated paleoclimatic rhythms of the late Pleistocene (that of the Mikulino interglaciation, lower Valdai interstades, and middle Valdai megainterstade). For instance, a poor substantiation of the Likhvin interglacial horizon gave rise to the unreasonable idea of lowering the stratigraphic level of the uppermost moraine down to the middle Pleistocene (Valueva *et al.*, 1969). However, the documented observations indicate that this moraine is stratigraphically above the Mikulino and lower Valdai interstadial deposits, whereas in the Rybinsk section, the Dnieper (middle Pleistocene) moraine is at the level of the third and fourth (from the surface) moraines.

This work is aimed to refine the Pleistocene stratigraphy and to outline periodic changes in the paleogeography of the region by means of new data and correlation schemes for reference sections in the Rybinsk–Yaroslavl region. Principal attention was focused on the stratification of the middle Pleistocene moraines in order to reconstruct landscapes and climatic environments during the Mikulino and early Valdai times, and to substantiate the viewpoint suggesting the progradation of the Kalinin glaciation into the Rybinsk–Tutaev area adjacent to the Volga River. Presented data on the geological position of the upper moraine in the Chermenino–Rybinsk (Fig. 2), Yakovka, and Dolgopolka reference sections are especially interesting in the context of recent discussion on the extreme front of the Kalinin moraine in the Russian plane.

In general, the glacial and glaciofluvial relief of the region is inherited from the Mesozoic denudation plane, whose surface is usually at the mean elevation of 60–80 m, though in river valleys it subsides down to 40–50 m, and is below sea level in local incisions now buried under sediments (for instance, near the Rybinsk the incision surface is at the level of –51 m). Irregularities of this pre-glacial surface were responsible for the dissection of the glacier terminus into fans and snouts (Fig. 1). The glacial relief is of a complex morphology, when median masses are regularly combined with terminal forms of different age and rank. At least six ter-

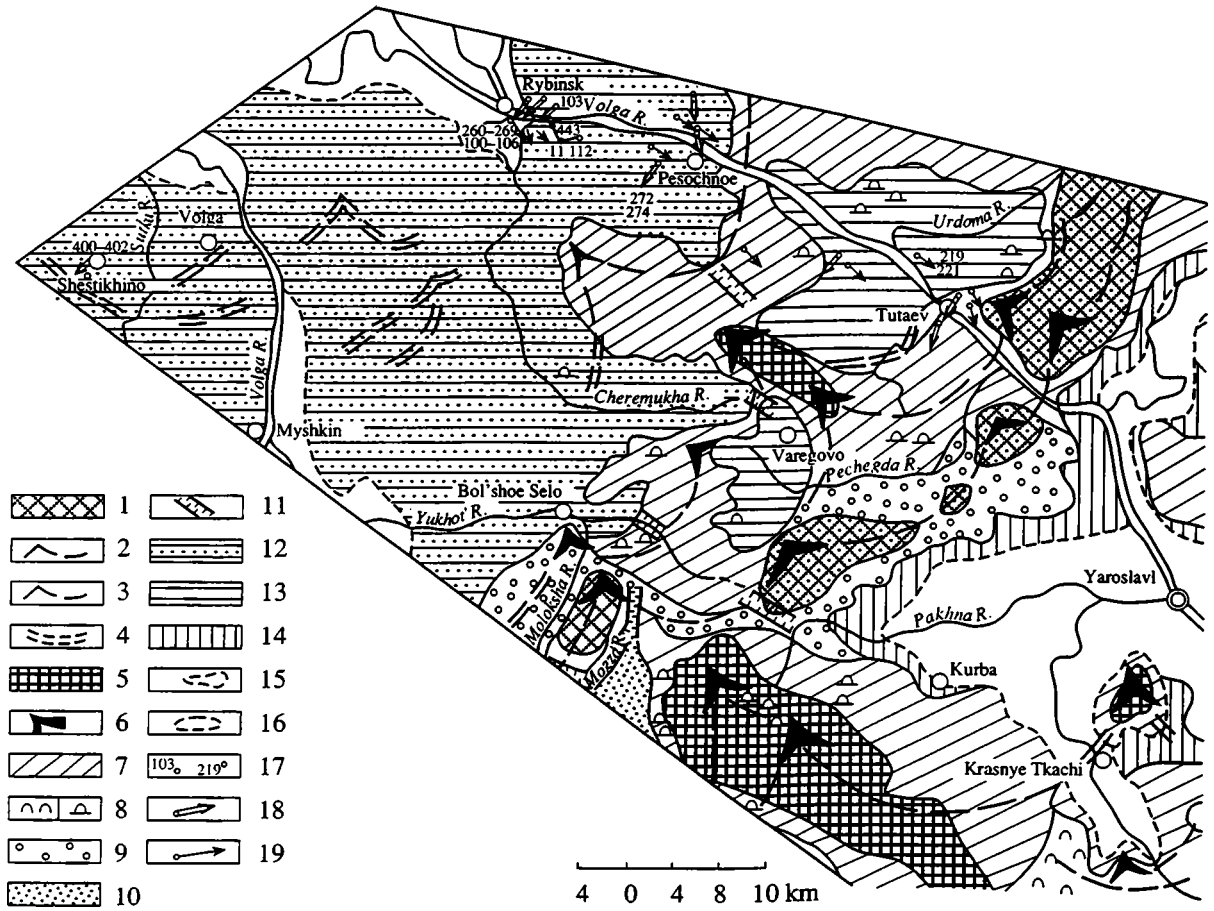

Fig. 1. Geomorphology of the Rybinsk–Yaroslavl region near the Volga River with elements of structure and dynamics of the glacial cover: *glacial topography*: (1) knob-and-kettle topography above the elevation of 140 m; (2) major zones of marginal forms; (3) minor zones of marginal forms; (4) chains of marginal forms in glaciofluvial plain; (5) interlobate knoll masses above the elevation of 150–160 m; (6) interlobate and intersnout corner masses; (7) interlobate undulating plain of major moraine; (8) knobs formed by passive ice: kame topography (a) and separate kames (b); *glaciofluvial topography*: (9) outwash plain around the terminus; (10) large radial valley sandar; (11) sandwashes; (12) glaciolacustrine plain (large glaciodepressions) below the elevation of 140 m; (13) dammed glaciofluvial basins; (14) glaciolacustrine terraces in the Yaroslavl and Rostov lowlands (areas below the elevation of 120 m); *lacustrine terraces*: (15) terraces below the elevation of 115 m in the Mologa–Sheksna lowland; (16) terraces below the elevation of 110 m in the Yaroslavl–Kostroma lowland; (17) wells and reference sections; (18) movement direction for the Dnieper glaciation (as implied by oriented inclusions in moraine); (19) ditto for the Moscow glaciation.

minal zones have been distinguished in the Rybinsk–Yaroslavl region adjacent to the Volga (Sudakova, 1990). The Kimry–Yaroslavl and Kalyazin–Danilov ridges (elev. up to 220 m) with accumulated glacial debris well exemplify such a terminal morphology. In the next, more distal zone of the terminal relief, which is confined to the glaciolacustrine depression crossing the Volga Valley near the village of Pesochnoe, glacial deposits represent a peculiar moraine facies produced by a minor snout from the greater Mologa–Sheksna depression.

Glaciolacustrine and moraine plains are more extensive. Below the elevation level of 140 m in the Mologa–Sheksna depression, we distinguished a series of glaciolacustrine and lacustrine terraces. Ancient accumulation levels corresponding to the ice-dammed glaciolacustrine basins of different ages are also detectable along both sides of the Volga Valley, where the third and fourth levels (elev. 25–40 m) are marked by the loess-like blanket deposits of a basin origin.

The second terrace, with relative elevation of 13–15 m, is well expressed in the Rybinsk–Tutaev interval of the valley. The loam–pebble–boulder horizon of thin (1–2 m) moraine is covered here with alluvium (2–4 m); it differs in composition from the relief-forming Moscow moraine, and is stratigraphically higher than the Mikulino beds.

The Rybinsk exposure, up to 35 m in height and 1600 m across, is on the right side of the Volga River. This exposure and supplementary data from the deep Well 110a (Valueva *et al.*, 1969; Novskii, 1975; *Razrezy* ..., 1977) are of primary importance for subdividing the glacial deposits and refining the moraine

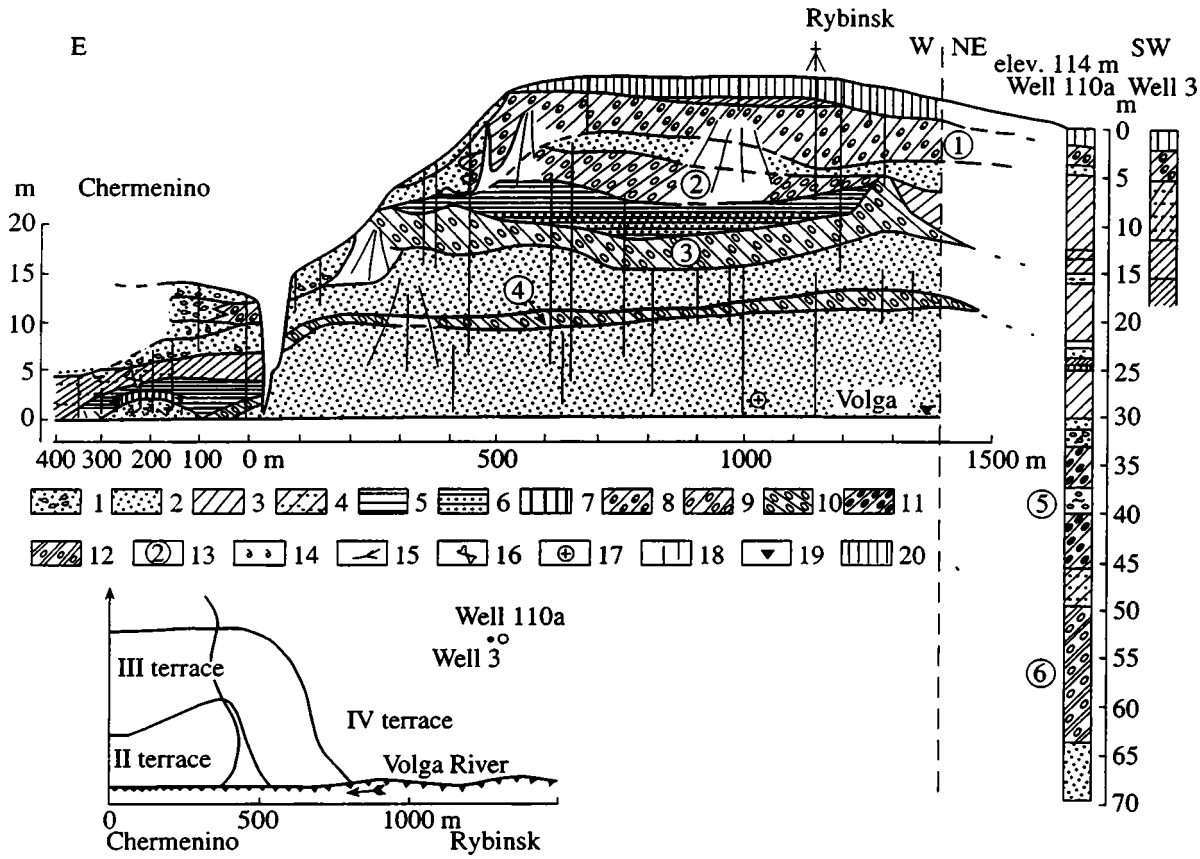

Fig. 2. Lithostratigraphy of Pleistocene deposits in the Rybinsk–Chermenino section, the right side of the Volga River.
(1) sand with pebbles; (2) sand; (3) loam; (4) sandy loam; (5) clay; (6) sandy clay; (7) humic sediments; (8) upper Pleistocene (Kalinin) moraine; (9) middle Pleistocene (Moscow) moraine; (10) middle Pleistocene (Dnieper) moraine; (11, 12) lower Pleistocene moraines; (13) numbers of moraine horizons; (14) molluscan shells; (15) plant remains; (16) localities with large mammal bones; (17) remains of small mammals; (18) stripped outcrops; (19) water level of the Volga River; (20) loess-like loam.

ages. Our field observations and results of laboratory investigations provided grounds to outline the general structure of the section (Fig. 2), and to elucidate the stratigraphic positions of separate glacial horizons; for instance, to detect stratigraphic relations of the middle Pleistocene moraine with older and younger deposits, and to identify the Kalinin moraine. We distinguished six moraine horizons within the range of this composite section, with four of them exposed in the Rybinsk locality above the Volga waterline. All of the horizons are different in color, lithologic features, mineralogy, and magnetic parameters.

According to the fauna of small mammals from sands at the base of the exposed section (*Razrezy ...*, 1977; determinations by A.K. Agadzhanyan) and recent thermoluminiscence dating (*Sudakova et al.*, 1992; measurements by L.T. Voskovskaya), the moraine, which is fourth from the surface, is not older than 230 ± 45 ka (TLM-408), i.e., corresponding in age to the Dnieper moraine. The main forms among mammals are lemmings of the genus *Dicrostonys*, the representatives of the tundra biocoenosis similar in their evolutionary level to lemmings from the loess-like deposits underlying moraine in the Chekalino section, and both units seem to be of the same age. Therefore, the beds, which are exposed in the Rybinsk locality near the Volga water level, should be referred to the middle Pleistocene. Two other moraine horizons have been recovered below this level by Well 110a located 300 m away from the outcrop, and they presumably are of the early Pleistocene age.

The former moraine is gray-brown in color, rich in sandy material, and contains up to 15% of clayey fraction. Its heavy mineral fraction is represented by garnet (20%), epidote (18%), and hornblende (15%) associated with glauconite (up to 1%), sulfide, disthene, and staurolite (up to 5% each). All the minerals are derived from the Jurassic and Cretaceous rocks. The clayey fraction consists of hydromica (40–50%), montmorillonite (up to 40%), and chlorite (about 20%); the latter is more abundant here than in higher moraine horizons. The hydromica supply in these deposits was from the Paleozoic deposits of a transit provenance, while local provenances with Mesozoic rocks were the source areas for montmorillonite and chlorite.

The third (from the surface) moraine is similar in its lithologic and magnetic (the reverse polarity) parameters to the fourth moraine (*Razrezy* ..., 1977). It also shows similar orientation of clastic material, suggesting that glaciers were moving toward the south-southwest (Karpukhin and Sudakova, 1972). This implies that both moraines were formed during the Dnieper interval. A member of clay-sandy deposits with palynological remains separates the third and overlying moraine horizons. The member is gray, and mostly consists of gravel-size clasts with a minor (about 5%) admixture of aleuritic material. Its association of heavy minerals is also of the garnet–epidote–hornblende type (with the relative abundance of the minerals equal to 15, 18, and 25% respectively). As compared to the higher moraines, this association is less rich in Fe-oxides and Fe-hydroxides (less than 10%). Clay minerals are represented here by hydromica (40–45%), montmorillonite (30–40%), chlorite, and kaolinite (these two minerals up to 20% in sum).

Everywhere in the Rybinsk-Yaroslavl region, including the Rybinsk section proper, the Dnieper moraine is distinct due to its dark brown, or gray color. As compared to the Moscow moraine, these deposits are more dense and display a lower sorting index; their composition is more closely related to underlying rocks. For instance, their mineral spectra are more rich in garnet and sulfides, which are very characteristic of the Mesozoic provenances, and simultaneously depleted in hornblende, the guide mineral of the Fennoscandia provenances. Another factor discriminating the Dnieper and Moscow moraines is a higher proportion (about 75%) of clasts derived from Paleozoic–Mesozoic sedimentary rocks in the Dnieper moraine, containing only 24% of other clast types. In addition, the Dnieper moraine is more rich in fragments of metamorphic rocks, while granite and quartz fragments are more abundant in the Moscow moraine. As it is recorded in the Chermenino, Pesochnoe, Sonokhta, Rykusha, Sel'tso, and Dolgopolka sections, the Dnieper moraine also shows a regular orientation of elongated clasts, suggesting that during the respective time, the glacier movements were directed toward the south and south-southwest (*Razrezy* ..., 1977).

By analogy with other sections in the Yaroslavl region, we referred the second moraine of the Rybinsk section to the Moscow horizon. The guide mineral association from this moraine is of the epidote–garnet–hornblende type, and discriminates this level from higher and lower glacial beds. The total amount of hornblende–pyroxene fraction is about 30%; the content of garnet and epidote is equal to 10 and 8%, respectively. Hydromica and montmorillonite are almost equally abundant in the clay fraction (about 40%), where they associate with chlorite and kaolinite (15% in sum). In the Rybinsk–Yaroslavl region adjacent to the Volga River, the relief-forming Moscow moraine is relatively thick (up to 30 m) and widespread in depressions and water-parting plains, extending to marginal areas of knob-and-kettle topography, the accumulation zones of terminal glacial facies. The Moscow moraine is usually colored in reddish tints; it is less dense, more rich in sand (up to 40%) and gravel (13% in average), but depleted in clay fraction (6–12%). Its mineral composition is distinct because of a higher proportion of hornblende. The debris composition at this level is different from that of the Dnieper moraine, since the igneous rock clasts are more abundant. According to Karpukhin's data, the indicator stones from the Moscow moraine are composed of the Ladoga rapakivi, Shoksha quartzite, and volcanic rocks of the Suisar complex, all of them typical of the provenance extending in Karelia from the Segozero to Medvezh'egorsk (Karpukhin and Sudakova, 1972). Sedimentary rocks (mainly carbonates) compose less than 70% of the erratic material, whose origin is consistent with the orientation of elongated blocks suggesting that the glacier was moving from the northwest toward the southeast. Thus, the different orientation and dissimilar composition of the erratic material in the Moscow and Dnieper moraines show that they are related to disparate glaciations and provenances. The mentioned diagnostic features are important for the regional correlation of the Moscow and Dnieper glaciohorizons all over the Yaroslavl' region adjacent to the Volga River.

The first moraine below the top of the Rybinsk section is distinct because of the reddish tint and high concentration of coarse rock fragments. The moraine loam is not very compact, being rich in aleuritic (up to 15%) and clayey (about 30%) material. Its mineral association of the garnet–epidote–hornblende type is enriched in Scandinavian components, for instance, in amphibole and pyroxene (30–35%). Epidote (up to 21%), along with Fe-oxides and Fe-hydroxides (up to 20% in sum), are also more abundant here as compared to the underlying moraine. The upper moraine displays the highest content of montmorillonite (40–50%) and hydromica (about 40%), while simultaneously depleted in chlorite and kaolinite (10% and lesser). The low chlorite content in the upper moraine is a result of the low resistance of this mineral against the weathering and reworking processes which affected the older moraines.

The upper Pleistocene deposits of the region also include the lacustrine sediments and blankets of loesslike material. Alluvial facies are significantly less abundant among them, because all large river valleys in this area are young, and during the early Valdai time, they were still at the stage of semifluvial lakes.

The Mikulino deposits are widespread in the Yaroslavl' region, and from the viewpoint of recent discussion on the Kalinin glaciaton terminus, they are the key beds elucidating the stratigraphic position of overlying rocks. In the second 15-m-high terrace, which is between villages of Chermenino and Yakovka on the right side of the Volga River, the lacustrine deposits of the Mikulino time are exposed for a distance of 400 m,

and near the first village they rest directly on the middle Pleistocene moraine (Fig. 2). The overlying beds consist of greenish gray, horizontally laminated lacustrine loam (2–3 m in thickness), which yielded the plevifemoral bone of *Coleodonta antiquitatis*. The loam beds with an erosion surface underlie the 4-m-thick member of inequigrained sands with scattered pebble and gravel inclusions; the top portion of the member includes an interlayer with abundant molluscan shells. Above the member, there is the reddish brown moraine (Fig. 2) extending as the continuous horizon from Chermenino to Yakovka; locally it is overfilled with boulders and blocks of the Fennoscandian rocks. The clear stratigraphic relations of this moraine with the lacustrine Mikulino beds is a good argument to refer it to the upper Pleistocene, and to correlate it in time with the Kalinin moraine.

The Chermenino section was studied for a long time (Arslanov *et al.*, 1967; Novskii, 1959; *Rel'ef* ..., 1961; Sukachev *et al.*, 1961), and the Mikulino age of its deposits, which is beyond doubt, is compatible with our data (Fig. 3). According to thermoluminiscence dating, sediments at the base of the Mikulino beds are 125 ± 10 ka old (TLM-289; Sudakova *et al.*, 1992). The relevant spore–pollen diagram is identical to diagrams characterizing the Mikulino deposits in the central Russian plain.

Bearing in mind the controversial age interpretation (Valueva *et al.*, 1969; Novskii, 1975; *Razrezy* ..., 1977; etc.) of reference beds in the 35-m-high terrace near Rybinsk (Fig. 2), we paid special attention to the lacustrine deposits here underlying the first red moraine, and to similar sediments which have been recovered by Wells 110a and 3 located 500 and 700 m away from the exposure near the village of Semenovskaya. The thick (26–30 m) recovered member consists of alternating dark gray clay and loam with plant detritus, felty peat, and small molluscan shells. As follows from Novskii's data (1975), Well 110a penetrated interlayers of black combustible shale and dark gray loam at a depth of 23 m. It is clear that these lacustrine sediments, which rest on the erosion surface of a moraine with a basal pebble bed and are rich in organic matter, were accumulated in a vast and deep basin. However, these beds are absent in coastal exposures of the Volga River, and their stratigraphic position was inadequately interpreted.

Valueva *et al.* (1969) assigned the depth interval of 22–30 m of Well 110a to the Likhvin interglacial (palynozones L_1–L_6); however, this conclusion seems to be insufficiently argued. According to published data, the interglacial deposits properly corresponding to the optimum were identified using a single sample from a very thin layer, where these researchers recorded a maximum (for the entire section) in the content of pollen grains representing such genera as *Quercus*, *Tilia*, *Fagus*, *Alnus*, and *Abies* (less than 25% in sum). In all other samples below and above the latter, the content of thermophilic elements does not exceed 10%. Exotic forms, usually diverse in the Likhvin deposits, are represented here only by *Tsuga* (two pollen grains) and by two species of *Osmunda*. The same exotic elements have also been distinguished in the Mikulino beds, for instance, in Exposure 514 of the Cheremushnik gully (*Putevoditel'*..., 1981). Serrate patterns in the diagram reflect the irregular distribution of certain elements of the spore–pollen spectrum at the depth interval of 23.0–23.5 m corresponding to the interglacial optimum, and respectively, they point to recurrent erosion and gaps in sedimentation of that time. These types of sediments (muddy loam saturated with organic matter of the sapropelic type) suggest that interglacial deposits accumulated within a coastal limnic zone very sensitive to the water level fluctuations. As for the initial and terminal phases, their spore–pollen characteristics are identical for all interglaciations, including those of the Holocene time. Relevant spectra include the upper and lower *Picea* peaks bordering an optimum interval and coupled with spore–pollen assemblages of the *Pinus–Betula*, or *Pinus–Picea–Betula* forests. Using these data, we believe that deposits, which were recovered by Well 110a from the depth interval of 22–30 m, better correspond to the Mikulino interglacial. This seems to be quite reasonable, because their hypsometric position in the well is approximately at the level of the Mikulino beds exposed above the second terrace base near the Chermenino.

Spore–pollen spectra from the depth level of Well 110a, which is above the Mikulino beds, are of the periglacial type. They correspond to the beginning of the early Valdai glaciation. The higher interval of 20.5–21.5 m, where *Picea* pollen is more abundant, shows a somewhat lower content of those periglacial elements, which were attributed by Valueva *et al.* (1969) to the interstadial flora. In terms of the spore–pollen spectra and geological features, these deposits are similar to those sediments of the Chermenino section, which correspond to the first early Valdai interstade according to data of Spiridonova *et al.* (1976). In our opinion, the unit recovered by the well is also correlative with this interstade.

The paleogeographic situation corresponding to the sedimentation period of overlying lacustrine deposits was studied by using a section of Well 3 near the village of Semenovskaya, which is correlative to that of Well 110a at the depth interval of 0–17.7 m. Two floral assemblages were distinguished in the respective spore–pollen diagram (Fig. 4).

The first of them exemplifies deposits from the depth interval of 10.0–17.7 m, where they consist of dark to greenish gray loam bearing plant remains. Pollen from this interval is rich in specimens of herbaceous plants (up to 60%), predominantly of xerophytes, such as *Chenopodium* (up to 40%), *Artemisia* (up to 60%), and *Ephedra* (rare grains). The pine (25–60%) and birch (20–50%) pollen is dominant among arboreal

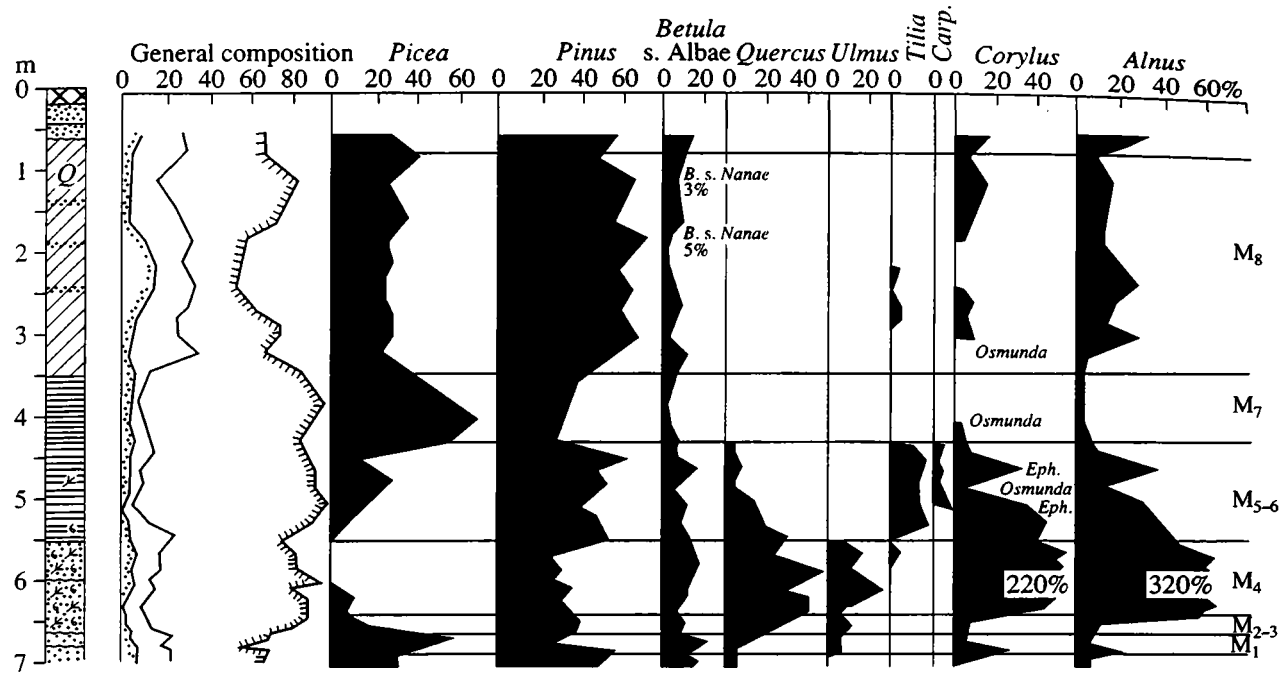

Fig. 3. Spore–pollen diagram for the Mikulino deposits near the village of Chermenino (symbols as in Fig. 4).

specimens, also including spruce (less than 20%), alder (below 10%), and dwarf birch (10–20%) pollen. Among the spores, there are dominant grains (50–98%) of true mosses associated with the constantly present specimens of *Sphagnum* and the accessory spores of *Selaginella selaginoides* and *Lycopodium selago*. These spore–pollen spectra are characteristic of the periglacial vegetation, which included the *Chenopodium–Artemisia* assemblages of open landscapes, and the bog plants typical of depressions. Rare forests of that time mostly consisted of pine and birch trees, with an admixture of spruce and alder in the moist areas. The vegetation was formed under the severe climate of the periglacial zones.

The second spore–pollen spectrum corresponds to the depth interval of 5.7–10.0 m, where deposits consist of dark gray loam with poorly petrified organic remains. It differs from the above spectrum in the much higher abundance of arboreal pollen (60–80%) and lower content of herbaceous specimens (10–20%). The proportion of spruce pollen grows upward (up to 59%) versus the decreasing content of dwarf and normal birch pollen. The pine pollen is not very abundant (40–50%), and accessory elements of the spectrum are represented by *Abies* cf. *sibirica*, some broad-leaved trees (*Tilia* and *Quercus*), and herbaceous (mostly mesophilous) plants, such as wormwood, goosefoot, and forbs. The spores include dominant species of true and bog mosses associated with ferns. Such spore–pollen spectra suggest relations to the dark coniferous forests of the northern taiga type. Climatic conditions were less severe than before, but more cold than at present, which is typical of an interstadial period (the second early Valdai interstade).

Thus, on the basis of the considered diagrams, we may suggest that during the period corresponding in Well 3 to the depth interval of 5.7–17.7 m, the vegetation near the glaciation terminus was replaced by forest assemblages of the second early Valdai interstade. Climatic fluctuations distinguishable in drilled sections (wells 3 and 110a) well fit the general evolutionary trend of the Valdai epoch. Even the most favorable periods of the middle Valdai time were not warmer than at present (*Chetvertichnaya ...*, 1984), and our analysis shows that the studied lacustrine deposits should be referred to the early Valdai time. The detected stratigraphic relations for the natural exposures and Well 110a are shown in Fig 2; they suggest the Kalinin age for the moraine overlying the lacustrine deposits, and one can see that there is no reliable biostratigraphic ground to assign an older age to this horizon.

The Kalinin moraine forms the uppermost bed (1–4 m in thickness) of the second terrace, extending from the village of Chermenino to the mouth of the Yakovka Creek. It rests as a blanket horizon on the periglacial deposits overlying the Mikulino lacustrine beds, and consists of a reddish brown loam with large and abundant boulders of variable composition exposed in outcrops of the Yakovka quarry. The erratic blocks are represented by the following rock types: gray and red granites (about 32%, including 1% of rapakivi); quartzite and Shoksha sandstone (6%); intermediate to basic igneous rocks (11% in sum); gneisses and granite-gneisses (4%); and schists (1%). The accumulated

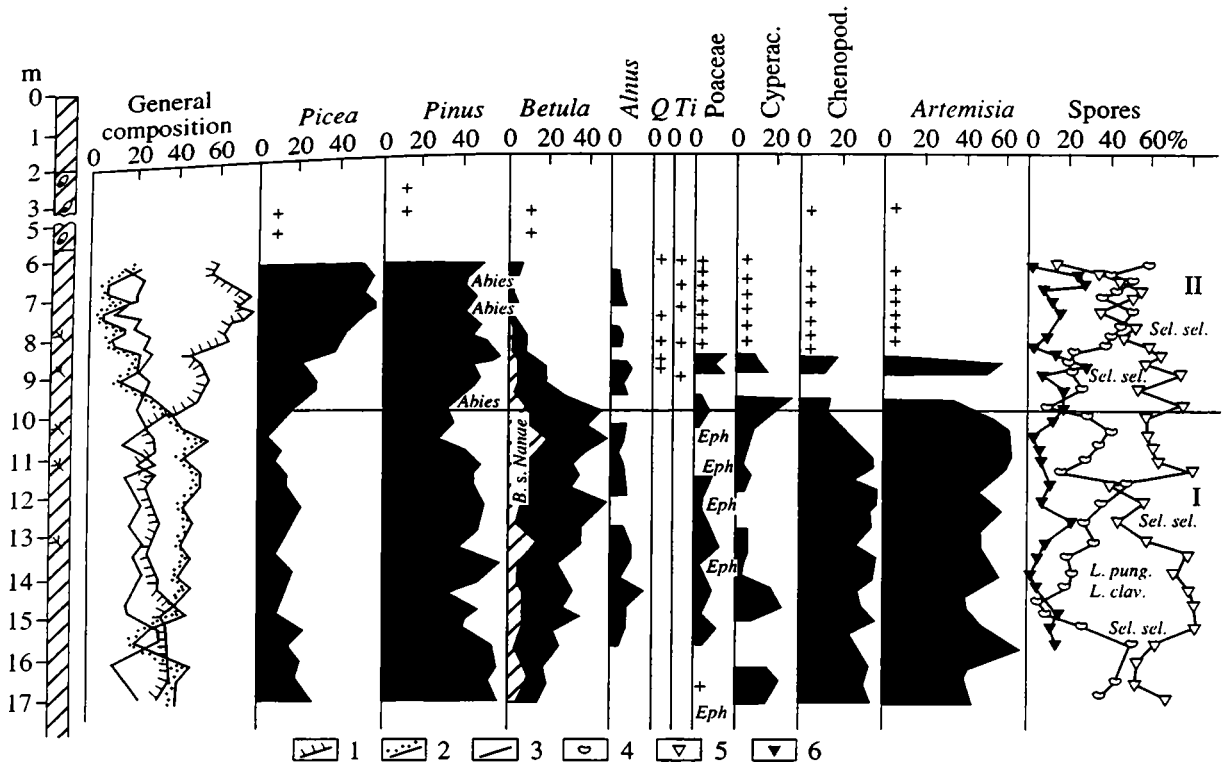

Fig. 4. Spore–pollen diagram for deposits from the Well 3.
1—pollen of arboreal plants; 2—pollen of herbaceous plants; 3—spores; 4—ferns; 5—true mosses; 6—bog mosses.

boulders are frequently more than 1 m in diameter. According to samples from the Yakovka quarry, the mineral composition of the Kalinin moraine is similar to that of the Moscow moraine in the Rybinsk region. The guide association here is of the epidote–garnet–hornblende type with the following proportions of minerals: dominant exotic hornblende—up to 25%; garnet—13%; epidote—10%; other minerals as accessory. Similar to the upper moraine in the Rybinsk section, this one also includes Fe-oxides and Fe-hydroxides (about 15% in sum); it is free of sulfides, but persistently shows the admixture of glauconite (up to 5%), which is assimilated from the Cretaceous rocks of a transit provenance. The respective group of clay minerals includes montmorillonite and hydromica (each up to 40–45%) with an admixture of kaolinite and chlorite (10–15% in sum). The moraine under consideration lies on the Mikulino sediments, whose age interval in the Dolgopolka section is, according to thermoluminiscence dating (Sudakova *et al.*, 1992), from 107 ± 11 ka (TLM-241) to 151 ± 24 ka (TLM-242). The above data suggest that the migration of the Kalinin glacial snout along the Volga valley was quite probable.

The representative Dolgopolka section is also of a reference rank for refining the stratigraphy of the middle–upper Valdai deposits. The second terrace of this section definitely comprises the middle Valdai sediments (*Razrezy...*, 1977). Here, as in the Grazhdanskii Prospect section (Kashin, the Sukhona River), the beds with a radiocarbon dating of 41 290 ± 320 years (LU-181) and pollen spectra of the northern taiga type (Arslanov *et al.*, 1974) correspond to the period of respective climatic optimum. The overlying periglacial deposits with large boulders were formed under the colder conditions of the late Valdai glaciation, when its terminus approached the Mologa–Sheksna depression. The terrace section also demonstrates features, which point to repeated water level changes in the past lake basin (Bazilevskaya *et al.*, 1988).

The loess accumulation in the Yaroslavl' region near the Volga River was closely related to the lake system evolution in this area. The distribution of loessal loams demonstrate their geomorphologic confinement to fluctuated water levels (Sudakova and Bazilevskaya, 1976). Obvious cycles in the loess blanket structure (*Razrezy ...*, 1977) reflect the periodic character of respective sedimentation, which was especially intense during the second half of the late Pleistocene, as it is evident from recent thermoluminiscence dating (Sudakova *et al.*, 1992).

In general, our study suggests the following conclusions:

(1) In the Rybinsk reference section (Fig. 2), the Dnieper age of the fourth moraine below the top is well substantiated by bio- and lithostratigraphic data. Consequently, the higher moraine horizons are of the Mos-

cow and Kalinin ages. The moraine beds, which have been recovered by drilling below the Volga water level, should be referred to the early Pleistocene glaciation.

(2) The palynological study of lacustrine deposits occurring at the terrace base near the village of Chermenino and corresponding to the entire interval of the Mikulino interstade reveals here classical palynozones M_1–M_8.

(3) We distinguished traces of two optimums and one transition to the cooling period in the post-Mikulino lacustrine deposits. Detailed palynological data suggest that these sediments accumulated in a deep basin near the glaciation terminus, which twice retreated backward, leaving behind interstadial conditions. The second interstadial warming was more mild, and resulted in a wider distribution of the dark coniferous forests of the northern taiga type over the Yaroslavl' region near the Volga River.

(4) In the Rybinsk–Tutaev area, the late Pleistocene age of the moraine, which is persistently exposed in the valley sections, is detectable from its stratigraphic position above the Mikulino beds. In addition, the Dolgopolka section shows that this moraine of the Kalinin level is overlain by dated deposits of the middle Valdai optimum.

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Evolution of the biosphere

Age of the Shand Gol Fauna and Evolution of Central Asia Mammals in the Oligocene

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Abstract—Analysis of the Oligocene mammalian fauna of Central Asia with consideration of the data on climatic evolution and geochronology revealed relations between major biotic events, which occurred in the studied region, and global events, and elucidated peculiarities of prochoreses and faunal exchange. The Shand Gol fauna belonged to the Central Asia biochore, whose mammal community remained almost stable throughout the Oligocene. It experienced minor changes because of the evolution of the autochtonous species and immigration of strangers during the regression. The appearance and extinction of the Shand Gol fauna and three successive stages of its evolution are related to global events across the Eocene–Oligocene and Oligocene–Miocene boundaries.

Key words: *Mammal fauna, Oligocene, Central Asia, evolution, biochronology.*

Studies on the Shand Gol mammalian fauna of Mongolia are crucial for understanding the processes which occurred at the beginning of the late Cenozoic in Central Asia. The Lake Valley, the main area of the Shand Gol fauna distribution, is one of the promising areas for investigation of faunal evolution during the second half of the Paleogene and beginning of the Neogene. Biotic events that occurred in the area are radiometrically dated, and can be correlated with similar events in distant regions. The Lake Valley area is also important because some recent mammalian families, which appeared in Central Asia at the Eocene–Oligocene boundary, evolved in this very area owing to the formation of a peculiar landscape—climatic and geobotanical environments. The knowledge of adaptation and distribution patterns of these early taxa representatives serves a basis for the correlation of biotic events in Central Asia with those in distant regions.

Mammalian remains in the Lake Valley were first found at the beginning of the 1920s during the field work of the Central Asia expedition of the American Museum of Natural History. Investigations here were performed in the late 1940s by the Mongolian paleontological expedition of the USSR Academy of Sciences headed by I.A. Efremov, and subsequently, since 1969, by the joint Soviet–Mongolian paleontological expedition. Results of the research of the materials obtained by these expeditions served a basis for this paper.

The studies were carried out in connection with the International meeting on UNESCO project 326 *The Oligocene–Miocene Transition in the Northern Hemisphere*, which was held in Ulan Bator in August, 1993. Particular attention was paid to ruminants. The representatives of this quickly evolved group are good indicators of the climatic and landscape changes.

Most of the fossil remains were sampled in the Shand Gol Formation composed of variegated predominantly sandy-clayey deposits of mainly proluvial-diluvial origin intercalated with basalt lava flows. These deposits form the reference horizon in the Cenozoic stratigraphic scale of Mongolia. Radiometric datings of basalt layers show that the base of the Shand Gol Formation is older than 32 Ma, and its top layers are younger than 24 Ma but older than 20 Ma. The latter dating indicates the age of basalts at the base of the Lo Formation. This formation consists of lacustrine and alluvial deposits, and contains the remains of early Miocene mammals in its basal beds.

Until recently, the mammalian fauna from the basal and top layers of the Shand Gol Formation was considered to be uniform. Initially, it was assigned to the Miocene (Berkey and Ganger, 1923). Later, the fauna was referred to different stratigraphic ranges such as the lower Oligocene (Matthew and Granger, 1923, 1924b), Aquitainian (Teilhard and Leroy, 1942), middle Oligocene (Flerov, 1961; Flerov and Yanovskaya,

1971; Dashzeveg, 1971; Belyaeva *et al.*, 1974; Savage and Russel, 1983), middle Oligocene (Mellett, 1968), upper Oligocene (Simpson, 1947), and middle–upper Oligocene (Badamgarav *et al.*, 1975; Devyatkin, 1981, 1993).

The evolutionary lineages *Tsaganomys altaicus–Beatomus bisus* of cylindrodontid (Shevyreva, 1974, 1976) and *Eumeryx culminis–E. imbellis–Eumeryx* sp. of cervids (Vislobokova, 1983, 1990a) discovered in the Shand Gol fauna indicate that the latter evolved during a long period. It was shown that the initial stage of this fauna development corresponds to the early Oligocene, whereas the terminal stage correlates with the late Oligocene faunas (Vislobokova, 1990a).

Analysis of published data and fossil remains from 13 localities in the Lake Valley, and depressions of Cisaltai, Transaltai Gobi, and Middle Altai Gobi showed that the appearance and extinction of the Shand Gol fauna are related to global events at the Eocene–Oligocene and Oligocene–Miocene boundaries. The faunal changes, which evidently correspond to the major landscape–climatic reorganizations, are also recorded in the interval between these boundaries.

The establishment of several evolutionary levels in various mammalian groups and radiometric datings obtained from the basalt layers allowed the evolutionary stages in the Shand Gol fauna to be distinguished and correlated with the Paleogene and Neogene mammal biozones of Europe, and mammal ages of North America.

Further research will probably reveal more complete characteristics of the separate stages in the Shand Gol fauna evolution, although this may meet with some difficulties, as most of the collected fossil remains were sampled at the surface and, thus, their stratigraphic position is uncertain.

The results obtained allowed us to enlighten the major biotic events in Central Asia, which were marked by faunal renovation because of both the evolution of autochtonous taxa and the migrational events.

COMPOSITION OF THE SHAND GOL FAUNA AND ITS EVOLUTION

The studies performed by American, Polish, Mongolian, and Russian paleontologists resulted in a description of the Shand Gol assemblage, which comprises 64 species of 43 mammal genera from 19 families and eight orders (Matthew and Granger, 1923, 1924a, 1924b; Trofimov, 1957, 1958; Shevyreva, 1965, 1976; Dashzeveg, 1971; McKenna and Holton, 1967; Mellett, 1968; Kowalski, 1974; Sych, 1975; Vislobokova, 1983; Vislobokova *et al.*, in press). By now, the following forms are determined in the Shand Gol fauna:

Insectivora. Erinaceidae: *Palaeoscaptor acridens*, *Amphechinus rectus*, *Tupaiodon morrisi*, ?*T. minutus*, and *Exallerix shandgolensis*;

Lagomorpha. Ochotonidae: *Desmatolagus gobiensis*, *D. robustus*, *D. simplex*, *D. youngi*, *Ochotonolagus argyropuloi*, and *Sinolagomys tatalgolicus*; and Leporidae: ?*Gobiolagus teilhardi*, *Procaprolagus vetustus*, *P. mongolicus*, *P. orlovi*, and *P. maximus*;

Rodentia. Aplodontidae: ?*Plesispermophilus lohiculus* and ?*Prosciurus* sp.; Cylindrodontidae: *Pseudocylindrodon mongoliense*, *Tsaganomys altaicus*, *Cyclomylus lohensis*, and *Beatomus bisus*; Cricetidae: *Cricetops dormitor*, *C. elephantus*, *Eumys asiaticus*, and *Selenomys mimicus*; Rhizomyidae: *Tachyoryctoides tatalgolicus*; and Ctenodactylidae: *Tataromys sigmodon*, *T. plicidens*, *T. deflexus*, and *Karakoromys decessus* (=*Woodomys chelkaris*);

Mesonichia. Didymoconidae: *Didymoconus colgatei* and *D. berkeyi*;

Creodonta. Hyaenodontidae: *Hyaenodon pervagus*, *H. dubius* (=*H. aimardi*), *H. ambiguus*, and *H. compressus*;

Carnivora. Miacidae: ?*Viverravus constans*; Canidae: ?*Cyonodictis elegans* and *C. constans*; Ursidae: *Amphicynodon teilhardi*; Viverridae: *Amphicticeps shackelfordi* and *Palaeoprionodon gracilis*; Mustelidae: *Palaeogale ulysses*, *P. parvula*, and *Plesictis* sp.; and Felidae: *Nimravus* sp. and *Proailurus* sp.;

Perissodactyla. Indricotheriidae: *Indricotherium transuralicum* (=*I. grangeri*);

Artiodactyla. Gelocidae: *Lophiomeryx angarae*, *L. gobiae*, *L.* cf. *mouchelini*, *L.* cf. *chalaniati*, *Gobiomeryx dubius*, *Miomeryx* cf. *altaicus*, and *Pseudomeryx gobiensis*; Cervidae: *Eumeryx culminis*, *E. imbellis*, *E.* sp., *Amphitragulus* cf. *quercyi*, and *Dremotherium* cf. *guthi*; and Bovidae: *Palaeohypsodontus asiaticus*.

The distribution of these forms among the localities was given in earlier publications (Mellett, 1968; Devyatkin, 1981; Vislobokova *et al.*, in press).

Nine genera (*Desmatolagus*, ?*Plesispermophilus*, *Prosciurus*, *Pseudocylindrodon*, *Eumys*, *Hyaenodon*, *Viverravus*, *Nimravus*, and *Palaeogale*) were typical of the holarctic provinces. Three genera of small mammals (*Amphechinus*, *Desmatolagus*, and *Plesispermophilus*), almost all carnivora genera (except the *Viverravus* and *Amphicticeps*), and three ruminant genera (*Lophiomeryx*, *Amphitragulus*, and *Dremotherium*) were recorded in the European faunas. A considerable part of the species was, until recently, known only from Central Asia. Others were reported from faunas of Kazakhstan, China, and Europe.

Three successive stages can be distinguished in the Shand Gol fauna evolution (Table 1):

(1) *The early stage (34–32 Ma)* is represented by fauna from the basal layers of the Shand Gol Formation below the basalts with age values of 31–32, 32 ± 2.0, and 31 ± 2.0 Ma (the Tatal Gol, Shand Gol, and Ikhe Argaluntu Nuru localities, respectively) (Mellett, 1968; Devyatkin, 1981). These strata contain small primitive *Lophiomeryx angarae* and *L. gobiae*, primitive heloc-

Table 1. The Oligocene mammalian fauna of Central Asia

Localities				First appeared genera	Characteristic species
Mongolia		China	Kazakhstan		
Lo		Xiejia	Xiejia Lanzhou	Agyspe	
Shand Gol	early			*Argyromys, Aralomys, Palaeocastor, "Oioceros," Eotragus*	
				Amphilagus, Beatomus, Amphitragulus, Dremotherium	
		Yikebulage	Yikebulage, Qianlishan, beds 18–21, Taben Buluk, Shargaltein Tal	Zhairam	
				Tachyoryctoides, Sinolagomys	
	middle	Wulanbulage	Qianlishan, beds 12–17, Sanshengon (Saint Jacques) Shuitonkon (San Tan Kho) Ulantatal	Chelkar Teniz Kyzyl Zhak Myneskesuiek	
				Tsaganomys, Selenomys, Tataromys, Karakoromys, Allacerops, Pseudomeryx, Prodremotherium, Palaeohypsodontus	
Ergilyeen Dzo	late		Qianlishan, beds 1–12, Tsaganbulage	Kiin Kerish Aktau, base of the Aktau Formation	
				Procaprolagus, Eumys, Pseudocylindrodon, Paraceratherium, Aprotodon, Schizotherium, Gobiomeryx, Eumeryx	

Yagan Tologoi
Tatsin Gol
(<24 Ma)

Tatsin Gol
(>24 Ma),
Tsagan Obo

Shunkht
Shand Gol
Tatal Gol
(<32 Ma)

Tatal Gol
(>32 Ma)
Ikhe Argaluntu
Nuru
Menkhen Teg,
Ergilyeen Dzo

Desmatolagus ardynense
D. vetustus
D. gobiensis
D. robustus
Sinolagomys tatalgolicus
D. pusillus
S. kansuensis
S. pachygnathus
Tsaganomys altaicus
Beatomus bisus
Ardynomys
Cricetops minor
C. dormitor
Tachyoryctoides pachygnatus
T. kokonorensis
Tataromys sigmodon
T. suni
Indricotherium parvum
I. transouralicum
Lophiomeryx gobiae
L. angarae
L. cf. mouchelini
L. turgaicus
L. cf. chalaniati
Gobiomeryx dubius
Pseudomeryx gobiensis
Prodremotherium
E. imbellis
Eumeryx culminis
Eumeryx sp.
Amphitragulus cf. *quercyi*

ids (*Miomeryx* cf. *altaicus* and *Gobiomeryx dubius*) characteristic of the early Oligocene Ergilyeen Dzo (Ardyn Obo) fauna (Matthew and Granger, 1925), as well as the most primitive cervid species *Eumeryx culminis*, the descendant of the late Eocene *Archaeomeryx*. The following mammalian remains, probably, are also from these layers: leporids of the *Gobiolagus* genus known from the upper Eocene–lower Oligocene of Asia; cylindrodontids of the *Pseudocylindrodon* genus reported from the lower Oligocene deposits of North America; carnivora genera *Cynodictis* and *Palaeoprinodon* characteristic of the upper Eocene–lower Oligocene of Europe; and deltatheriids of the *Didymoconus* genus and species *Hyaenodon dubius* occurring at level 6 of the Quercy section.

The synchronous Ergilyeen Dzo mammalian fauna of southeastern Mongolia is correlative with the Chadronian (38–32 Ma) of North America because of the similarity of the Asian brontotheriid species *Parabrontops gobiensis* and American *Brontops brachycephalis* (Yanovskaya, 1981).

In China, this stage is correlative to the mammalian fauna from the Urtyn Obo, Tsaganbulage, and other formations (Li and Ting, 1983; Russel and Zhai, 1987), which correspond to the fauna of Wang's (1992) second horizon.

In Europe, the stage corresponds to the MP 21 Zone and level 6 in the Quercy section (Hoogbutsel) with *Desmatolagus* sp., *Plesispermophilus atavus*, *Hyaenodon dubius*, *Amphicynodon gracilis*, *Cynodictis palmidens*, and others (Remy *et al.*, 1987). The Paleogene mammal zones are dated using magnetostratigraphic data. The dating of 33.1 Ma is adopted for the MP 21 Zone. Its lower boundary corresponds to the Eocene–Oligocene boundary and correlates with the boundary between chrons C13R and C13N, the age of which is estimated to be 33.7 Ma (Leveque, 1993; Vianey-Liaud *et al.*, 1993).

In Kazakhstan, the fauna from the lower Aktau Formation in the Aktau Mountains (Dzhungar Alatau) and that from the lower Tuzkabak Formation in the Mt. Kiin Kerish area (Zaisan depression) are close to the early Shand Gol fauna.

(2) *The middle stage of the Shand Gol mammalian fauna (32–29.5 Ma)* is represented by remains occurring in the middle part of the Shand Gol Formation above the basalts 32 Ma old (the Shand Gol, Tatal Gol, and Shunkht localities). Characteristic of this stage are more evolved *Eumeryx imbellis*, and helocids *Lophiomeryx* cf. *mouchelini* and *Pseudomeryx gobiensis*. Most of the species distinguished in the Shand Gol fauna belong to this stage. The most abundant are lagomorphs (25%), among which the ochotonid species *Desmatolagus gobiensis*, cylindrodontid species *Tsaganomys altaicus* (13%), and cricetid form *Cricetops dormitor* (36%) are predominant (Mellett, 1968).

In Inner Mongolia, correlative faunas occur in the Ulantatal locality, where *Cyclomylus lohensis*, *Selenomys mimicus*, *Palaeogale ulysses*, *Eumeryx culminis*, *Palaeohypsodontus* cf. *asiaticus*, and *Hanhaicerus qii* and others were recorded (Huang, 1985, 1987), and in the top beds 12–17 of the Wulanbulage Formation in the Qianlishan Hills area, which yielded *Desmatolagus gobiensis*, *Tsaganomys*, *Tataromys sigmodon*, *Cricetops dormitor*, *Plesiosminthus*, *Paraceratherium*, *Aprotodon*, and others (Wang *et al.*, 1981; Wang, 1987). The Shuitongon and Ninsya faunas (the San Tan Kho and Qinshian localities of Inner Mongolia, respectively; Wang *et al.*, 1981) can probably be correlated with the middle stage of the Shand Gol fauna.

In Kazakhstan, evidently coeval assemblages are represented by the Chelkar Teniz fauna with *Plesiospermophilus lohicus*, *Cyclomylus turgaicus*, *Cricetops dormitor*, *Indricotherium transouralicum*, *Allacerops turgaicum*, *Lophiomeryx turgaicus*, *Prodremotherium flerowi* and others, and by Myneskesuiek and Kyzyl Kaka faunas (Flerov, 1938; Trofimov, 1957; Shevyreva, 1976; Tleuberdina and Rayushkina, 1993).

The presence of *Lophiomeryx* species similar to *Lophiomeryx mouchelini* allows the middle stage of the Shand Gol fauna to be correlated with the MP 22 to 23 mammalian zones of Europe. The MP 23 Zone corresponds to the Chron C12R, and is dated as 30.5 Ma old (Leveque, 1993).

In North America, the Orella and Uitni faunas (32–31 Ma and 31–28 Ma, respectively) also correspond to the middle stage of the Shand Gol fauna.

(3) *The late stage of the Shand Gol fauna (29.5–22 Ma)* is represented by mammalian remains from the top strata of the Shand Gol Formation incorporating the basalt layer of 24 ± 1.0 Ma old (the Tatsin Gol, Tsagan Obo, Khatan Khairkhan, and Yagan Tologoi localities; Devyatkin, 1981).

The first half of the stage is characterized by the large evolved gelocids, such as *Lophiomeryx chalaniati* reported from the MP 24 to 28 zones of Europe. *Tachyoryctoides tatalgolicus* and *Sinolagomys tatalgolicus* sampled in the Tsagan Obo locality also are typical of this stage (Badamgarav *et al.*, 1975).

Dremotheriins *Amphitragulus* cf. *quercyi* and *Dremotherium* cf. *guthi*, more evolved as compared to the *Eumeryx* forms and similar to those recorded in the European MP 28–30 zones, appear in the second half of the stage, and *Tsaganomys altaicus* is replaced here by *Beatomus bisus*. The MP 28 Zone is 24.5 Ma old (Leveque, 1993), whereas datings for the MP 30 Zone are 22.5 Ma (Leveque, 1993) and 24.5 Ma (Burbank *et al.*, 1992).

In Inner Mongolia, the fauna from the Yikebulage Formation in the Qianlishan Hills area is correlative with that of the third stage. It includes primitive *Sinolagomys kansuensis*, *S. major*, and *S. gracilis* (=*S. minor*), as well as *Tachyoryctoides kokonorensis* and *Tataromys suni* reported from the lower Mocene of China (Wang *et al.*, 1981).

In the Ganshu province, the third stage fauna of the Shand Gol mammalian assemblage correlates with the Taben Buluk and Shargaltein Tal faunas with *Beatomus bisus*, *Tachyoryctoides obrutschewi*, *T. intermedium*, *Indricotherium grangeri*, and others.

In North America, this stage is represented by the early Aracarian fauna (28–24 Ma). In Kazakhstan, the Zhairema fauna is close to that of the third stage (Tleuberdina and Rayushkina, 1993).

IMPACT OF GLOBAL CLIMATIC EVENTS ON THE MAMMALIAN FAUNA OF CENTRAL ASIA

The two most significant changes in biota during the late Paleogene–beginning of the Neogene correspond to two global events, which occurred at the Eocene–Oligocene and Oligocene–Miocene boundaries.

MAMMAL FAUNA CHANGES AT THE EOCENE–OLIGOCENE BOUNDARY

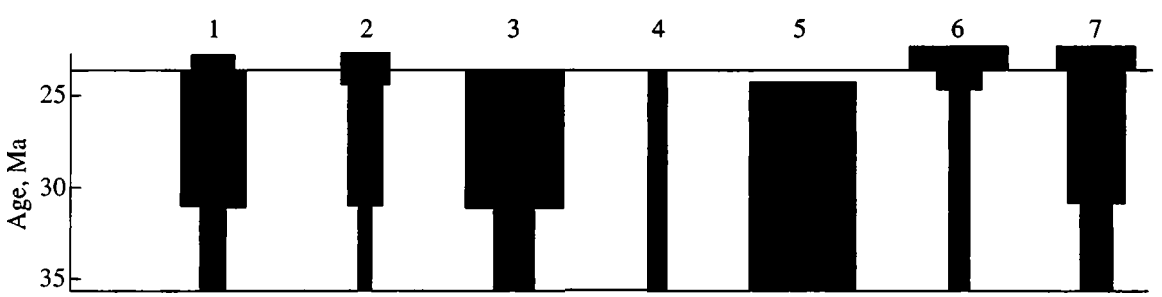
The first faunal renovation resulted from a global cooling in response to the Antarctic glaciation approximately 34 Ma ago. This renovation event is well pronounced in Europe (the so-called *Grand Coupure* event) and in North America. The replacement of archaic families with recent ones took place elsewhere at that time. The renovation rate of the mammalian fauna determined by the number of newly appeared genera per million years was 21 in Europe and 18 in North America (Savage and Russel, 1983). During this time interval, 78 and 96 new genera appeared in Europe and North America, respectively (Savage and Russel, 1983). Accelerated rates of mammal evolution resulted from their adaptation to new biotopes: the broad-leaved deciduous forests spread, according to Wolfe (1985), over areas of the 35°–50°N latitudes, and the savanna and semidesert biotopes occupied inner areas of these continents. The fauna renewal was also promoted by immigrants.

The sea level fall due to the Antarctic glaciation and regression of epicontinental seas resulted in the aridization of inner areas of Europe, formation of the global arid belt, and appearance of the peculiar Central Asian biochore. Beginning from the terminal Eocene, this belt became a zone of xerophytic vegetation, with abundant herbaceous coenoses and strongly thinned forests with dry areas of the savanna and semidesert type.

In parallel with the floral transformation, significant faunal changes occurred: mammalian forms had to adopt to the relatively coarse vegetational food, inhabit the hard-ground landscapes, and cross large distances. The first hypsodont shrews of the Cylindrodontidae family, as well as rhyzomyids and ctenodactylids, adopted to the arid and semiarid environments originated among the rodents. In addition, the deficiency of the vegetation biomass promoted the origination of ruminant forms with complicated ventricle fermentation of vegetative food that allowed the food resources to be utilized at a maximum.

Across the Eocene–Oligocene boundary, Central Asia became the origination center of herbivorous mammal groups such as ochotonids, indricotheriids, and higher ruminants (cervids and bovids), and the area of adaptive radiation of cylindrodontids, cricetids, and tragulids (figure). Not less than 20 genera appeared there. For instance, these are the first leporids of the *Procaprolagus* genus among the lagomorphs, cricetids of the *Cricetodon* and, probably, *Eumys* genera, as well as cylindrodontids of the *Cyclomylus* genus among the rodents. In the perissodactylid group, the Eocene dinoceros were replaced by chalicotheriids, indricotheriids, and rhinoceros that coexisted with brontotheriids and aminodontids. The primitive tapir-like lophialetids became extinct. The formation of new herbivorous communities and new habitat environments resulted in the transformation of carnivora communities. The Eocene creodonts and miacids gave way to representatives of recent Carnivora families: Canidae, Felidae, Viverridae, and Mustelidae.

The late Eocene Shara Murun fauna of Inner Mongolia (Ula Usu and other localities) with primitive tragulids of the *Archaeomeryx* genus was replaced by the Ergilyeen Dzo and Shand Gol faunas in Mongolia, and by correlative faunas with tragulids of the *Lophiomeryx*, *Miomeryx*, *Gobiomeryx*, and other genera in China.



Dynamics of generic composition of the background mammalian groups in Central Asia during the Oligocene.
1—Ochtonidae; 2—Cylindrodontidae; 3—Cricetidae; 4—Indricotheriidae; 5—Gelocidae; 6—Cervidae; 7—Bovidae.

Table 2. Faunal exchange between Central Asia, Europe, and North America in the Oligocene and at the beginning of the Miocene (arrows show the prochorese direction)

Land connections between North America and Europe at the Eocene–Oligocene boundary caused faunal exchange between Central Asia and these regions. The Beringian land bridge served as the migratory way for ochotonids (*Desmatolagus*), aplodontids (*Prosciurus*), cylindrodontids (*Pseudocylindrodon*), cricetids (*Eumys*), and mustelids (*Palaeogale*), although one cannot exclude that representatives of the latter genus migrated somewhat later (Table 2). Some of these animals (*Desmatolagus*, *Palaeogale*) may have dispersed along with the *Antracotherium* and *Entelodontus* species over western Europe. In addition, there were other forms common to Central Asia and Europe, such as aplodontids (*Plesiospermophilus*), cricetids (*Eucricetodon*), and carnivore species *Amphicynodon*, *Cynodictis*, and *Hyaenodon dubius* (Remy *et al.*, 1987).

MAMMAL FAUNA CHANGES DURING THE OLIGOCENE

The Oligocene was a time of further aridization of Central Asia with predominantly open landscapes. The further fall and fluctuations of average annual temperatures was the prevailing climatic feature. Throughout most of the Oligocene, inhabitants of the semi open and open sufficiently dry biotopes were widespread, whereas those of forest and wet biotopes were rare. The prevailing forms were ochotonids (*Desmatolagus*) and leporids (*Procaprolagus*) among lagomorphs, cylindrodontids (*Tsaganomys*), cricetids (*Cricetops*), rhyzomyids (*Tachyoryctoides*), and ctenodactylids (*Tataromys* and *Karakoromys*) among rodents, giant rhinoceros (*Indricotherium*) and chalicotheriids (*Schizotherium*) among perissodactylids, and small ruminants among artiodactylids. The latter group included the first hypsodont forms.

During approximately 12×10^6 years, the faunal evolution was gradual without drastic changes. This is evident from the changes in the genera composition and evolutionary level of some forms. In separate mammalian lineages, evolution caused the development of hypsodonty features, lengthening of limb distal sections, and reduction of side sections of the limbs.

At the beginning of the Oligocene, the Central Asian fauna evolved in sufficiently insulated environments, because the sea level rise resulted in flooding over the existing land bridges. Still, later, insignificant faunal exchange occurred repeatedly.

Insignificant faunal changes in Central Asia are recorded in the mid-early and mid-late Oligocene. They evidently resulted from increased climatic aridity in the region owing to regressions corresponding to three phases of the Antarctic ice shield growth. According to isotope data on foraminiferal shells from deep-sea sediments, they occurred 31–29, 25–24, and 23–22 Ma ago, respectively (Shackleton *et al.*, 1984; Shackleton, 1986; Keigwin and Keller, 1984; Miller *et al.*, 1985). Latitudinal migrations of mammals mainly in the east–west direction took place approximately at that time.

Approximately 32 Ma ago, western Europe was invaded by *Lophiomeryx*, felids *Nimravus*, and mustelids *Palaeogale*, which are recorded at levels 7 and 8 in the Quercy section (Remy *et al.*, 1987). In Central Asia, cylindrodontids *Tsaganomys*, cricetids *Cricetops*, and giant rhinoceros *Indricotherium* were widespread. In addition, cricetids *Selenomys*, ctenodactylids *Tataromys* and *Karakoromys*, rhinoceros forms *Allacerops* and *Aprotodon*, tragulids *Pseudomeryx* and *Prodremotherium*, small bovids *Palaeohypsodontus* and *Hanhaicerus* appeared for the first time, whereas *Brontotherium* became extinct. *Desmatolagus ardynense* and *D. vetustus* were replaced by *D. gobiensis* and *D. robustus* (Huang, 1987); *Lophiomeryx angarae* and *L. gobiae* were substituted by *L.* cf. *mouchelini*; *Indricotherium transouralicum* gave way to *I. parvum*, and *Eumeryx culminis* replaced *E. imbellus*.

In the mid-early Oligocene, Asian and European faunas were, evidently, separated from each other; however 30 Ma ago, the connection between these regions was restored again. The areas occupied by Tethys and Paratethys considerably decreased at that time. The arid belt widened northward and eastward covering the Turgai depression territory. In the south, it extended to Ordos.

In Central Asia at that time, a noticeable share of the mammalian fauna consisted of ochotonids *Sinolagomys*, rhizomyids *Tachyoryctoides*, and ctenodactylids *Tataromys*. *Lophiomeryx* cf. *mouchelini* was replaced by *L.* cf. *chalaniati* and more evolved *Eumeryx* sp. substituted *E. imbellis*. The *Desmatolagus* genus was represented by the species *D. pusillus* (Huang, 1987). *Sinolagomys tatalgolicus* was replaced by *S. kansuensis*, *S. major*, and *S. gracilis*, whereas *Tataromys suni* gave way to *T. sigmodon*. Approximately 30 Ma ago, *L.* cf. *chalaniati* and *Schizotherium* forms immigrated into Europe.

In the second half of the late Oligocene, lagomorphs *Amphilagus* together with cervids *Amphitragulus* and *Dremotherium* appeared in Central Asia, whereas gelocids, which survived in other areas, became extinct. *Beatomus bisus* replaced *Tsaganomys altaicus*. Approximately 25 Ma ago, Western Europe was invaded by *Hyaenodon compressus*, mustelids of the *Plesictis* genus, *Prodremotherium*, *Amphitragulus*, and *Dremotherium*. *Indricotherium* species simultaneously reached the Transylvania and Balkan regions (Codrea, 1989).

MAMMALIAN FAUNA CHANGES AT THE OLIGOCENE–MIOCENE BOUNDARY

The second major faunal renovation in Central Asia and gradual destruction of the biochore, which was formed earlier, related to landscape–climatic reorganizations at the Oligocene–Miocene transition that, in

turn, were caused by the humidization of climate in response to major transgression and orogeny. In the Lake Valley area, the sedimentation patterns sharply changed then: predominantly proluvial-diluvial Oligocene deposits were substituted at the beginning of the Miocene by lacustrine and alluvial facies. In Central Asia, the biotope variety increased, and forest inhabitants became more abundant.

As in Europe, this renovation most likely occurred approximately 22 Ma ago. It affected both the autochtonous species and immigrants from North America and Africa.

The renovation rate of mammalian faunas in Europe was somewhat lower at that time than across the Eocene–Oligocene boundary, whereas in North America, it was comparable with that of the Chadronian time (Savage and Russel, 1983). At the Oligocene–Miocene transition, 37 and 67 new genera appeared in Europe and North America, respectively (Savage and Russel, 1983).

At the Oligocene–Miocene boundary, groups which existed in Central Asia during the first half of the Paleogene became extinct, whereas representatives of the late Cenozoic families became more diverse. Extinct forms included creodonts, giant rhinoceros, aminodonts, entelodonts, and primitive tapiroids, whereas rhizomyids *Aralomys*, beavers of the *Palaeocastor* and *Capatanka* genera, and bovids *Oioceros* and *Eotragus* appeared for the first time (Sokolov, 1952). Some former genera were represented by more evolved species: *Sinolagomys kansuensis* was replaced by *S. pachygnathus*, *Amphilagus antiquus* by *A. ulmensis*, *Eucricetodon caducus* and *E. asiaticus* by *E. youngi*, and *Amphitragulus* cf. *quercyi* by *A. boulangeri*.

Using the Beringian land bridge, which existed 23–22 Ma ago, the *Palaeocastor* and *Capatanka* species migrated from North America to Asia and, on the contrary, the *Dremotherium* forms migrated from Asia to America.

In Central Asia, faunas of the Shand Gol type were substituted by assemblages occurring in the basal layers of the Lo and Oshi formations of Mongolia, and in basal strata of the Aral Formation of Kazakhstan (Agyspe, Shokysu and other localities) (Bendukidze, 1993; Tleuberdina and Rayushkina, 1993; and others), and also by faunas of the Xiejia stage in China (Qiu, 1990).

Renovation is observed not only in the terrestrial, but also in marine faunas. Changes are recorded among cetaceans in the Eastern Paratethys: archaeocets and squalodonts were replaced here by numerous dolphins, acrodolphins, and misticets (Mchedlidze, 1990).

CONCLUSION

(1) The Shand Gol mammalian fauna existed in Central Asia throughout the Oligocene. Its appearance and extinction are related to global events across the Eocene–Oligocene and Oligocene–Miocene boundaries that resulted in sharp changes in the landscape–climatic environments.

(2) In the Shand Gol fauna evolution, three successive stages, each characterized by peculiar evolutionary levels of small mammals and ruminants, can be defined: the early stage (34–32 Ma) with *Lophiomeryx angarae*, *L. gobiae*, and *Eumeryx culminis*; the middle stage (32–29.5 Ma) with *L.* cf. *mouchelini* and *Eumeryx imbellis*; and the late stage (29.5–22 Ma), the first half of which is marked by *L.* cf. *chalaniati* and *Eumeryx* sp., and the second half, by *Amphitragulus* cf. *quercyi* and *Dremotherium* cf. *guthi*.

(3) The Shand Gol fauna exemplifies the Central Asian biochore, the formation of which began in the terminal Eocene. In this biochore, accelerated evolution of mainly herbivore forms, which were adopted to grinding relatively coarse vegetative food, proceeded against the background of increasing aridization and formation of herbaceous coenoses and xerophytic vegetation. The major part of the mammalian community was represented by forms that appeared in this region at the Eocene–Oligocene boundary: ochotonids, cylindrodontids, cricetids, giant rhinoceros, diverse tragulids, and primitive bovids and cervids, which coexisted with representatives of archaic groups.

(4) The mammalian community of the Central Asian biochore was almost stable throughout the Oligocene. It experienced small changes at the species and genus levels because of evolution within separate lineages and immigration. The area of the Central Asian biochore changed with time, and depended on global events: in the earliest Oligocene, it was confined to the inner areas, though by the mid-early Oligocene, it was extending from the Turgai depression to Ordos.

(5) The appearance of land bridges during regressions and the extention of the Central Asian biochore promoted migration of some of its inhabitants to Europe, probably, via the Elburs area, Asia Minor, and the Balkans or along northern coast of the Paratethys. Western Europe was invaded by lagomorphs *Desmatolagus*, aplodontids *Plesispermophilus*, cricetids *Eucricetodon*, carnivores of the *Amphicynodon* genus, *Antracotherium*, and *Entelodon* during the first migratiory event (the MP 20–MP 21 boundary); by *Lophiomeryx*, felids *Nimravus*, and mustelids *Palaeogale* during the second migratiory event (zones MP 22–23); and first by gelocids *L.* cf. *chalaniati* and chalicatheriids *Schizotherium* and later by mustelids *Plesictis* and cervids of the *Amphitragulus* and *Dremotherium* genera during the third migratiory wave (MP 24–28 zones).

(6) Faunal exchange with North America occurred at the Eocene–Oligocene boundary, in the middle and terminal Oligocene, and at the beginning of the Miocene. During the first exchange event, *Desmatolagus*, *Prosciurus*, *Pseudocylindrondon*, *Eumys*, and *Palaeogale* crossed the Beringian land bridge. Felids *Nimravus* probably traveled this way at the beginning

of the Oligocene. At the beginning of the Aquitainian, cervids (*Dremotherium* and *Dremomericina*) migrated from Asia to North America.

(7) At the beginning of the Miocene, the mammalian community of the Central Asian biochore became extinct. One part of its constituents (ctenodactylids, paraceratheriids, tragulids, and others) survived as relicts in refugees of China, Kazakhstan, and Asia Minor. The other part, mainly represented by autochtonous small herbivore mammals (lagomorphs, cervids, and bovids), experienced intense adaptive radiation.

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The Stratigraphy and Small Mammals from the Aral Formation, the Altynshokysu Locality, Northern Aral Area

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Abstract—Two assemblages of small mammals were identified in the Lower Miocene rocks of the Altynshokysu locality where they are confined to different stratigraphic units. The results promote more thorough correlations within the Aral Formation and comparisons with other localities of mammals in Asia and mammalian biozones in Europe. A list of small mammals of the Aral fauna is given with the description of three new species—two of cricetid rodents and one of insectivores.

Key words: *Aral Formation, lower Miocene, Northern Aral area, stratigraphy, correlation, small mammals.*

INTRODUCTION

Small mammals, primarily insectivores, lagomorphs, and rodents, constitute one of the most representative groups for biostratigraphic correlations of continental Cenozoic rocks. This group is characterized by a rapid evolution of some of the taxa of small mammals, by their relative abundance in biocoenoses, and by a high sensitivity and prompt reaction to environmental changes. The remains of these animals are now more frequently used for the subdivision and correlation of continental rocks, including, in particular, the Paleogene–Neogene formations. The mammal biozones in Europe are as detailed as the zones based on marine microorganisms (Mein, 1979; Agusti *et al.*, 1988).

The Aral faunal assemblage also called paracerathe-rian, is insufficiently studied. Apart from the classic Akespe (Agyspe) locality discovered in 1930, only a few other localities can be compared with it in terms of the abundance of material; e.g., the Kuzhasai, Zhilansai, and Sayaken localities on the northwestern coast of the Aral Sea. There is information about the finds of mammalian remains in other parts of the Aral Formation. The age of the complex was debated for a long time (late Oligocene–early Miocene). Now it is referred to the Aquitanian.

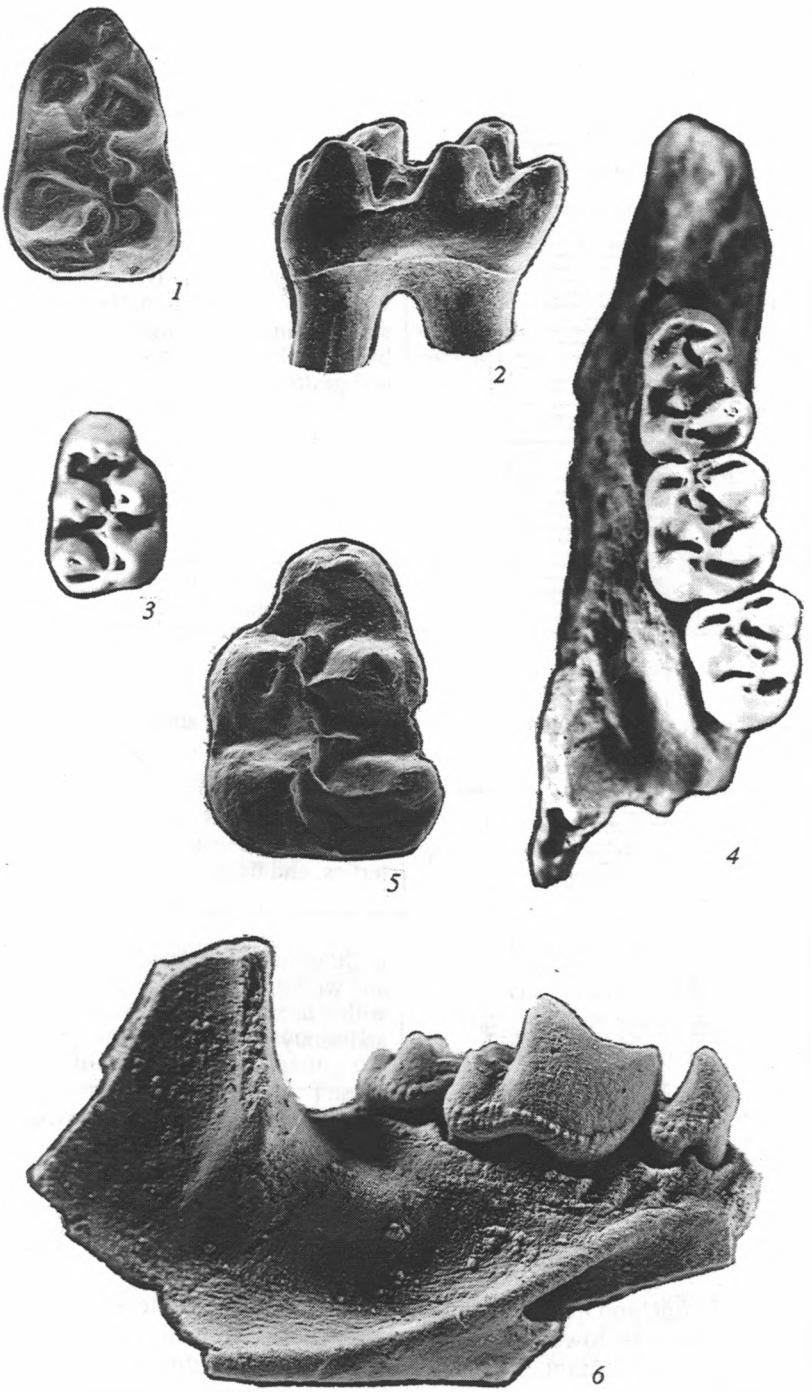
The Altynshokysu (Altyn–Chokusu) locality is 4 km northeast of the Chokusu Station and 45 km north of the Aral Sea coast line. The locality is unique due to its extensive taxonomic diversity of fossil vertebrates (various fishes, amphibia, reptiles, birds, mammals), and by the presence of marine organism remains (selachians, cetaceans). At the Altynshokysu, the Aral Formation is represented, apparently, in its complete volume, shows variable lithology, and has a maximal thickness (up to 20 m).

The material was obtained in 1992–1993 during my research in western Kazakhstan as a member of the expeditions of the Paleontological Institute of the Russian Academy of Sciences (PIN RAS). The mammalian remains were collected from the washed out surface or recovered by means of excavation and washing. In all, more than 500 different remains were collected and studied.

STRATIGRAPHY

The area north of the Aral Sea comprises numerous Paleogene and Miocene deposits represented, in most cases, by terrigenous–carbonate rocks. According to the environments of their formation, they are subdivided into "deposits of marine Paleogene" (Tas Aran, Saksaul'skaya and Chegan formations), the fresh and brackish-water deposits of the "Turgai Group" (Kutanbulak, Chilikty and Chagrai formations), and the brackish-water deposits of the Aral Formation (Yanshin, 1953; Ovechkin, 1962; Vyalov, 1964).

The time of the Aral Formation, also known as the "*Corbula* beds" or "beds with *Corbula helmerseni*," corresponds to the regressive phase of Paratethys, and implies lagoon and lacustrine sedimentation on a coastal plain. The age of the beds is now defined as the terminal Oligocene–early Miocene (Popov *et al.*, 1993). The Aral Formation occurs in the northern part of the Aral area, in eastern Ustyurt, and in the southern part of the Turgai plain; as a rule, it is represented by grayish green carbonate clays with intercalations of multicolor clays, silts, and sands. The Aral Formation varies laterally and in thickness. Normally, two types of sections are exposed: the western sandy–silty–clayey type with a small amount of carbonate rocks (Sayaken, Kyzhasai, and Zhilansai), and the eastern clayey–carbonate type with a small share of sandy and silty rocks (*Geologiya SSSR*, 1970). The Altynshokysu section obviously belongs to the eastern type, though the proportion of silts and sands in this locality is larger than that, for example, in the Akespe locality.



The early Miocene mammals from Altynshokysu locality (the Northern Aral area), Aral Formation.

1—Eumyarion tremulus sp. nov., no. 4516-70, holotype, right M_1, occlusal view, ×22; 2—same, labial view, ×22; *3—E. tremulus* sp. nov., no. 4516-60, right M^1; ×12; *4—Eucricetodon occasionalis* sp. nov., no. 4516-196, holotype; the left fragment of the lower jaw with M_1–M_3; ×12; *5—E. occasionalis* sp. nov., no. 4516-198, left M^1; ×22; *6—Exallerix efialtes* sp. nov., no. 4516-1, holotype, the right fragment of the lower jaw with P_4–M_2; ×5.

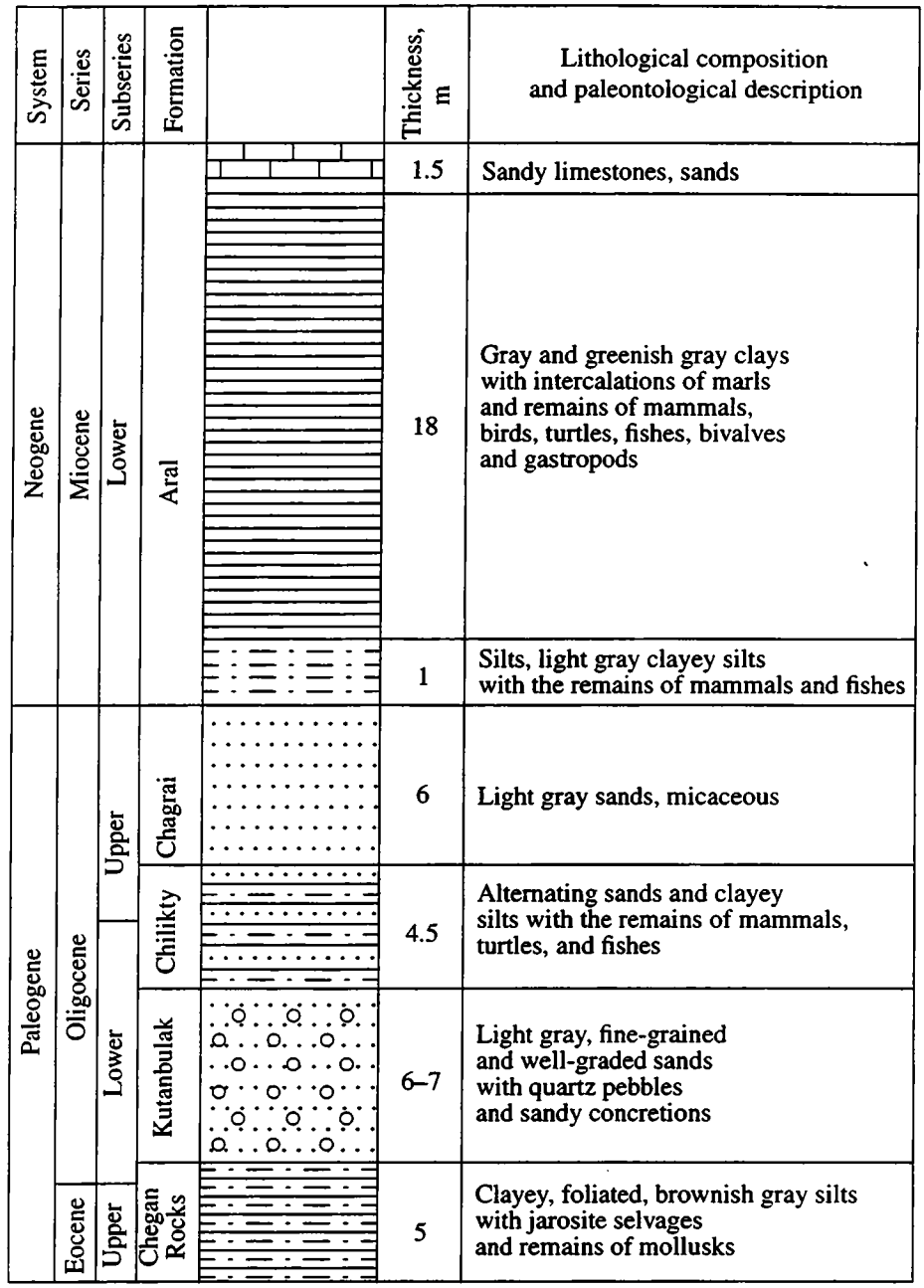


Fig. 1. The stratigraphic scheme of the Paleogene–Neogene deposits on the southwestern slope of the Altynshokysu Plateau.

The Altynshokysu plateau is flat, and its steep southwestern slope exposes the Eocene–lower Miocene deposits about 40 m thick (Chegan, Kutanbulak, Chilikty, Chagrai, and Aral formations; Fig. 1).

The Aral Formation concordantly overlies sands of the Chagrai Formation. The sands are fine-grained, light gray, horizontally laminated, well-graded, micaceous, and often kaolinitized; they contain numerous ferro-manganese concretions and interbeds, and also quartz and argillite pebbles up to 4 cm long.

The Aral Formation is exposed in the upper part of the slope, where it is 15–20 m thick and consists of clayey and carbonate–clayey rocks with a small admixture of clastic material. The formation is divisible into members with distinct lithological features. From the base upsection these are:

(1) Light gray clayey silts (1 m).

(2) Light gray and greenish gray clays with intercalations of marls (up to 18 m).

(3) Sandy limestones, sandstones, and sands (up to 1 m).

In addition to the typical shells of bivalve mollusks *Corbula helmerseni* Mikh. and gastropod remains, the clays and silts of the Aral Formation contain bones of

various fishes, amphibia, turtles, birds, and mammals (insectivores, lagomorphs, rodents, carnivores, cetaceans, perissodactyls, artiodactyls).

On the southwestern slope of the Altynshokysu plateau, the following members are exposed from the base upsection:

Member 1

Bed 1. Clayey silts, light bluish gray, sand-rich in an upper portion enclosing the bone-bearing lenses and intercalations. The main bone-bearing level is at the top of the bed. Fragments of bones are usually mixed with small pebbles and gravels (1 m).

The first bone-bearing level (I) contains the remains of the vertebrates listed below.

Fishes: *Odontaspis* sp., *Esox aralensis* Sytch., *Acipenser* sp., and diverse specimens of Percidae and Cyprinidae (preliminary definitions by Sychevskaya).

Amphibia: Cryptobranchiidae indet. (definition by Shishkin).

Mammals represented according to my determinations by the following groups: insectivores *Amphechinus* cf. *minimus* (Bohlin, 1942), Erinaceidae indet., *Proscapanus* sp., Desmaninae indet., *Gobisorex* aff. *kingae* Sulimski, 1970; lagomorphs *Desmatolagus simplex* (Argyropulo, 1940), *D.* aff. *shargaltensis* Bohlin, 1937; rodents *Steneofiber kumbulakensis* (Lytshev, 1970), *Plesiosminthus* sp., *Aralocricetodon schokensis* Bendukidze, 1993, *Eumyarion tremulus* sp. nov., *Eucricetodon occasionalis* sp. nov.; cetaceans ?Delphinoidea indet.; and artiodactyls *Amphitragulus* sp.

Member 2

Bed 2. Bright gray-green rather dense clays: they form the main gently inclined part of the slope and contain numerous shells of *Corbula helmerseni* Mikh., which often compose lenses, intercalations and partings (up to 15 cm thick) of white coquina consisting of detritus, fragments and whole shells of corbulas. There are quartz pebbles up to 3 cm long, and a rose-gray marl interlayer up to 10 cm thick at the top of the bed (2 m).

The second bone-bearing level (II) contains fish bones, mainly vertebrae of Percidae.

Bed 3. Clays, greenish gray with a lighter shade than those in Bed 2 (1.5 m).

Bed 4. Clays, yellowish gray; when weathered yellowish brown (0.5 m).

Bed 5. Carbonate clays, rather dense and fresh, dark gray, when weathered greenish gray, fragmented. At the top, there is an intercalation (7 cm) of rose-gray silty marl (1.5 m).

Bed 6. Light gray clays, yellowish gray after weathering. One meter above the bottom of the bed, there is the third bone-bearing level (III) with fish skeletons of Clupeidae indet. as defined by Sychevskaya (2.5 m).

Bed 7. A sequence of alternating gray (yellowish gray if weathered) and greenish gray clays (3 m).

Bed 8. Greenish gray clays, plastic if fresh, fragmented if weathered; they contain numerous operculas of freshwater gastropods *Bithynia adornatus* Tolst. in association with rare shells of *Corbula helmerseni* and gastropods, fragments of turtle carapaces, birds bones of Anatidae, Ergilornithidae, Charadrii (definition by Karkhu), and teeth and bones of mammals (4 m).

The fourth bone-bearing level (IV) yielded the following mammalian remains: insectivores *Exallerix efialtes* sp. nov., *Amphechinus* cf. *minimus* (Bohlin, 1942), *Mygalea lavrovi* Bendukidze, 1993, *Asthenoscapter* sp., *Gobisorex* aff. *kingae* Sulimski, 1970; lagomorphs *Desmatolagus* aff. *robustus* Matthew et Granger, 1923, *D.* aff. *shargaltensis* Bohlin, 1937, *Sinolagomys pachygnathus* Li et Qiu, 1980; rodents *Ansomys* sp., *Steneofiber kumbulakensis* (Lytshev, 1970), *Asiacastor* aff. *orientalis* Lytshev, 1988, *Eucricetodon occasionalis* sp. nov., *Aralomys gigas* Argyropulo, 1939, *Tachyoryctoides glikmani* Vorontzov, 1963, *Yindirtemys birgeri* Bendukidze, 1993; carnivores ?Stenoplesictinae indet., Mustelidae indet.; perissodactyls Rhinocerotoidea indet.; and artiodactyls *Amphitragulus* sp..

Member 3

Bed 9. Sands: light gray, fine-grained, silty, ferruginated (1m).

Bed 10. Sandy limestones: rose-gray, platy, containing rare casts of *Corbula helmerseni* and rounded fragments of mammalian bones. They are exposed at the armored plateau surface (up to 0.5 m).

THE PROBLEMS OF CORRELATION USING MAMMALS

The study of small mammals of the Aral fauna is obstructed by fragmentation and poor preservation of their remains. The description of most of their forms can be done only in an open nomenclature. A list of small mammals of the Aral assemblage and their localities is given in Table 1.

As a rule, two subassemblages of mammal species are recognized in the Aral fauna. The earliest was recovered at the Akespe locality, where it includes endemic rodents (*Eumysodon* and *Argyromys*), and lagomorphs *Desmatolagus simplex*. The Akespe level correlates either with the MN 1 zone of Mein's scale (Mein, 1979), or with MN 2 (Gabuniya, 1986) and, therefore, corresponds to the lower Aquitanian. The second subassemblage is typical of the Sayaken, Kuzhasai, and Zhilansai localities on the northwestern coast of the Aral Sea. These localities contain lagomorphs of the genus *Sinolagomys*, and rodents *Tachyoryctoides* and *Yindirtemys*. They are dated as the upper Aquitanian forms (Bendukidze, 1977). The ancient representatives of lagomorphs *Desmatolagus* aff. *shargaltensis*, beavers *Steneofiber kumbulakensis* and

Table 1. List of small mammals from the Aral faunal assemblages (after Argiropulo, 1939, 1940; Belyaeva and Borisyak, 1948; Bendukidze, 1977, 1989, 1993; Lavrov *et al.*, 1985; Gabuniya, 1981; Russel and Zhai, 1987; the finds of the author are marked with a cross in parenthesis)

Taxon	Locality							
	Early Aralian fauna		Late Aralian fauna					
	Akespe	Altynshokysu I	Altynshokysu IV	Shokysu	Akotau	Sayaken	Kuzhasai	Zhilansai
INSECTIVORA								
Erinaceidae:								
Echinosoricinae:								
Lantanotherium sp.	–	–	–	+	+	–	–	–
Erinaceinae:								
Exallerix efialtes sp. nov.	–	–	(+)	–	–	–	–	–
Amphechinus cf. *rectus* (Matthew et Granger, 1924)	–	–	–	+	–	–	–	+
A. minimus (Bohlin, 1942) (? =*Palaeoscaptor acridens* Matt. et Gr., 1924–Belyaeva, Borisyak, 1948)	+	–	–	+	–	–	–	–
A. cf. *minimus* (Bohlin, 1942)	–	(+)	(+)	–	–	–	–	–
Talpidae:								
Desmaninae:								
Mygalea lavrovi Bendukidze, 1993	–	–	(+)	+	–	+	–	–
Uropsilinae:								
Asthenoscapter sp.	–	–	(+)	+	–	–	–	–
Proscapanus sp.	–	(+)	–	+	–	–	–	–
Soricidae, Heterosoricinae:								
Gobisorex kingae Sulimski, 1970	–	–	–	+	–	–	–	–
G. aff. *kingae* Sulimski, 1970	–	(+)	(+)	+	+	–	–	–
LAGOMORPHA								
Palaeolagidae, Desmatolaginae:								
Desmatolagus aff. *robustus* Matt. et Gr., 1923	–	–	(+)	+	–	–	–	–
D. aff. *shargaltensis* Bohlin, 1937	+	(+)	(+)	+	+	+	–	–
D. simplex (Argyropulo, 1940) (=*D.* aff. *gobiensis* Matt. et Gr., 1923–Bendukidze, 1993)	+	(+)	–	+	–	–	–	–
Ochotonidae, Sinolagomyinae:								
Sinolagomys pachygnathus Li et Qiu, 1980 (=*S.* aff. *kansuensis* Bohlin, 1937; =*S.* aff. *gracilis* Bohlin, 1946–Bendukidze, 1993)	–	–	(+)	+	+	+	–	+
RODENTIA								
Aplodontidae, Ansomyinae:								
Ansomys sp.	–	–	(+)	–	–	–	–	–
Sciuridae:								
Palaeosciurus sp.	–	–	–	–	–	–	–	+
Dipodidae:								
Argyromys aralensis (Argyropulo, 1939) (=*Schaubeumys aralensis*)	+ (+)	–	–	–	–	–	–	–
A. woodi (Argyropulo, 1939)	+	–	–	–	–	–	–	–
?*Protalactaga borissiaki* Argyropulo, 1939	+	–	–	–	–	–	–	–
Zapodidae:								
Plesiosminthus sp.	–	(+)	–	–	–	–	–	–

Table 1. (Contd.)

Taxon	Locality							
	Early Aralian fauna		Late Aralian fauna					
	Akespe	Altynshokysu I	Altynshokysu IV	Shokysu	Akotau	Sayaken	Kuzhasai	Zhilansai
Parasminthus aff. *tangingoli* Bohlin, 1946	–	–	–	+	–	–	–	–
Cricetidae, Eucricetodontinae:								
Eucricetodon sajakensis Bendukidze, 1993 (=*E.* aff. *asiaticus* (Matt. et Gr., 1923); =*E.* aff. *deploratus* (Shevyreva, 1967)	–	–	–	+	+	+	–	–
E. occasionalis sp. nov. (=*E.* aff. *caducus* (Shevyreva, 1967)–Bendukidze, 1993)	+	(+)	(+)	+	+	–	–	–
Eumyarion tremulus sp. nov. (=*Eucricetodon* aff. *youngi* Li et Qiu, 1980–Bendukidze, 1993)	(+)	(+)	–	+	–	–	–	–
Aralocricetodon schokensis Bendukidze, 1993	–	(+)	–	+	–	–	–	–
Ctenodactylidae:								
Yindirtemys sajakensis Bendukidze, 1993 (=*Tataromys* aff. *deflexus* Teilhard, 1926–Bendukidze, 1977)	–	–	–	+	+	+	–	–
Y. birgeri Bendukidze, 1993	–	–	(+)	+	+	+	–	–
"Tataromys" cf. *sigmodon* Matt. et Gr., 1923	–	–	–	–	–	–	+	+
Rhizomyidae:								
Tachyoryctoides aff. *obrutschewi* Bohlin, 1946	+	–	–	+	+	+	–	–
T. glikmani (Vorontzov, 1963) (=*T. spurius* (Argyr., 1939)–Bendukidze, 1993)	–	–	(+)	+	–	–	–	–
Aralomys gigas Argyropulo, 1939	+	–	(+)	+	+	+	–	–
Aralomys sp.	–	–	–	+	+	+	+	–
Eumysodon spurius Argyropulo, 1939	+	–	–	–	–	–	–	–
E. orlovi Argyropulo, 1939	+	–	–	–	–	–	–	–
Castoridae: Trogontheriinae:								
Steneofiber kumbulakensis (Lytshev, 1970) (=*Propalaeocastor kumbulakensis*)	+ (+)	(+)	(+)	+	–	–	–	–
Castoroidinae:								
Palaeocastor sp. (=*Agnotocastor* sp.)	+	–	–	+	–	–	–	–
Capatanka schokensis Bendukidze, 1993	+	–	–	+	+	–	–	–
C. aff. *schokensis* Bend., 1993	–	–	–	–	–	–	–	+
Capacikala sajakensis Bendukidze, 1993	–	–	–	–	–	+	–	–
C. aff. *sajakensis* Bend., 1993	–	–	–	+	+	–	–	–
C. cf. *sciuroides* (Matt., 1907)	–	–	–	+	+	–	–	–
Asiacastorinae:								
Asiacastor aff. *orientalis* Lytshev, 1988	–	–	(+)	+	+	–	–	–

Palaeocastor sp., cricetid rodents of genus *Eucricetodon* and *Aralomys gigas* occur in both subassemblages (Table 1).

The lowermost bone-bearing level (I) at the Altynshokysu locality is of the early Aquitanian age according to the remains of small mammals. The comparison of *Eucricetodon occasionalis* sp. nov. with the European counterpart of the genus shows that by the degree of morphological evolution of teeth, the Aralian representative is similar to the stage of *E. herperius* Engesser, 1985 (zone MN 1 of the Mein's scale, the Paulhiac locality; Engesser, 1985). *E. occasionalis* sp. nov. is probably the direct descendant of the Oligocene *E. caducus* (Shevyreva, 1967), which is also referred to

the lineage *E. praecursor* (Schaub, 1925)—*E. collatus* (Schaub, 1925)—*E. hesperius* (stage of *E. praecursor*; Shevyreva, 1967). *Eumyarion tremulus* sp. nov. is similar to *E. carbonicus* Bruijn et Sarac, 1991, from the lower Miocene in Turkey (zone MN 1 or 2, the Harami I locality); the former, however, differs from the latter by the reduced posterior arms of hypoconids that is typical of the Miocene *Eumyarion* from Europe. The structure of the teeth of *Plesiosminthus* sp. from the first bone-bearing level at the Altynshokysu locality is rather similar to that of *P. myarion* Schaub, 1930 (the double protoloph M^2 and the well-developed posterior arm of protoconid M_2), the form typical of the lower Miocene in Europe (zone MN 1–2a).

Eumyarion tremulus sp. nov. and *Desmatolagus simplex* were also found in the Akespe level. Therefore, we may compare this and the lower bone-bearing level of the Altynshokysu locality, and consider the latter as representing the early Aral subassemblage. According to results of the analysis of the composition of small mammals, this subassemblage can be dated as corresponding to zone MN 1.

The remains of *Sinolagomys pachygnathus* Li et Qiu, 1980, recovered from the upper (IV) bone-bearing level of the Altynshokysu locality, suggests its Aquitanian age. *S. pachygnatus* is recovered and described from the Xiejia Formation (Qinghai, China) dated by zones MN 2–4 (Li *et al.*, 1981), MN 1–2 (Li *et al.*, 1984), or MN 1–3 (Qiu and Qiu, 1990). The remains of Desmatolaginae forms, defined as *Desmatolagus* aff. *robustus* and *D.* aff. *shargaltensis*, suggest a somewhat older age of this assemblage. However, in fact, these lagomorphs are represented by forms with some progressive features. For example, *D.* aff. *robustus* is larger than *D. robustus* Matthew et Granger, 1923, and has cement in the anteroexternal fold of P_3 and in the folds separating the hypoconulids from the talonids of P_4–M_2. *D.* aff. *shargaltensis* shows an obvious tendency to merging of roots of P_4–M_2 and to reduction of hypoconulids. It seems evident that these two forms are independent species.

Ansomys sp. from the Altynshokysu seems to be, as it is evident from the P_4 structure, an intermediate form between the late Oligocene *A. shantungensis* (Rensberger et Li, 1986) and the middle Miocene *A. orientalis* Qiu, 1987 and *A. shanwangensis* Qiu et Sun, 1988 (China). A fairly isolated large hypoconid and the undeveloped low crest of metastylid imply the close affinity of *Ansomys* sp. to *A. shantungensis*, whereas the well developed mezostylid, the accessory crests of the talonid basin, ectolophid, and hypolophid show its resemblance to *A. orientalis* (features typical of the Miocene Ansomyinae; Qiu et Sun, 1988). Perhaps, *Aralomys* should be treated as a separate Miocene genus of Rhizomyidae (Bendukidze, 1993) in so far as it differs from the ancestral Oligocene *Tachyoryctoides* by a somewhat reduced M_3. *Tachyoryctoides kokonorensis* Li et Qiu, 1980, the representative of rhizomyids from Xiejia, may also belong to this genus. The ctenodactylids *Yindirtemys birgeri* and, in particular, *Y. sajakensis* are very similar to *Y. suni* Li et Qiu, 1980, from the same locality. The beavers *Steneofiber kumbulakensis* and *Asiacastor orientalis* also occur in the Akzhar Formation of the Zaisan depression dated by the early Miocene (Lychev, 1987).

At present, we know *Mygalea* and *Asthenoscapter* only from the Miocene. Unlike *E. hsandagolensis* McKenna et Holton, 1967, of the Oligocene age, *Exallerix efialtes* sp. nov. has certain features progressive for Brachyericini taxa; i.e., the relatively larger and narrower M_1 with a cutting paralophid, and reduced metaconid, small P_4, and massive masseteric ridge. On the other hand, *Amphechinus* cf. *minimus* and *Gobisorex* aff. *kingae* are almost identical to corresponding species of Oligocene insectivores of Mongolia in the size and structure of teeth (Sulimski, 1970).

The listed species belong to the late Aralian subassemblage of small mammals of the Aral fauna (Table 1). Their age, apparently, corresponds to the Xiejian Age of mammals of the Chinese scale (MN 2–3).

Bendukidze (1993) described the abundant complex of small mammals from the Akotau and Shokysu localities of the studied region. He identified two bone-bearing levels, and compared the lower assemblage with Akespe and the upper one with Sayaken, Kuzhasai, and Zhilansai faunas. As described, they correspond to beds 8 and 9 of the Altynshokysu section, and, in my opinion, both are of the late Aralian age. Their typical species are found in the Sayaken, but not in the Akespe section and the lower (I) bone-bearing level at the Altynshokysu locality (Table 1).

If these conclusions are confirmed by further research, then the Altynshkysu section can be considered as the most important locality of the Aral faunal complex comprising remains of small mammals of both its early and late stages.

A comparison of the Aral assemblages with the Asiatic and European Oligocene–Miocene faunas shows that they contain rather numerous forms that retained the features of late Oligocene organization; this well-known feature also characterizes the Aquitanian mammals of Europe. This is typical primarily of erinaceids, paleolagids, aplodontids, ctenodactylids, certain rhizomyids and cricetids. The greatest similarity is noted between these and the late Oligocene faunas of China (Shargaltein Gol and Taben Buluk localities). On the other hand, it is obvious that many of the above taxa have specific features distinguishing them from their Oligocene ancestors. Some of the species are identical or very similar to the species from the Xiejia fauna (China), which at the moment is the most ancient Miocene fauna of Asia; e.g., by the morphological evolution of their teeth, the rodents *Eucricetodon* and *Eumyarion*, are comparable with the European and western Asiatic (Turkey) species typical of the Aquitanian.

In sum, it seems obvious that before the beginning of the early Miocene, the Aral fauna was at the last stage of the Oligocene epoch of evolution of the mammalian fauna in Central Asia.

PALEOECOLOGY

The mammals of the Aral fauna dwelt in different biotopes of two main groups; i.e., in the humid forests near rivers and lakes, and in dry open and semiopen landscapes. This conclusion is based on the results of a study of fossil remains of small mammals from the Akespe and Altynshokysu localities, confirming that the Aral fauna included two main ecological groups, among which the dwellers of open and semiopen areas were dominant (Flerov and Yanovskaya, 1971; *Zoogeografiya* ..., 1974).

The dwellers of open landscapes (meadows, dense bushes, steppe or savannalike areas, dry watershed woods, and open woodland) were hedgehogs *Amphechinus* and *Exallerix*, lagomorphs *Desmatolagus* and *Sinolagomys*, rhizomyid representatives of *Aralomys* and *Tachyoryctoides*, and *Yindirtemys* of the ctenodactylid family. The ancient cricetids lived in moderately humid and dry woods on watershed divides, and were the ecological analogues to modern mice (Shevyreva, 1967). According to the morphology of teeth, *Aralocricetodon schokensis* could feed on cellular tissue, whereas *Eumyarion tremulus* and *Eucricetodon occasionalis* preferred protein nutriment. *Ansomys* was probably dwelling in woods and bush thickets, while the diverse beavers, muskrats, and soricids inhabited areas near rivers and lakes.

Large mammals also comprise two ecological assemblages. The animals living near water, such as *Aprotodon borissiaki* and Aminodontidae, constitute one of the assemblages; the other includes obvious dewellers of open landscapes; i.e., gigantic rhinoceros *Paraceratherium prohorovi* and *Aceratherium aralense*, hyracodonts, prodremotheres, and others (Yanshin, 1953; Lavrov, 1959; Lychev and Aubekerova, 1971; Bendukidze, 1979).

The identified groups of biocoenosis correlate well with palynological data characteristic of extensive areas covered with deciduous and coniferous woods with subtropical elements and xerophytic vegetation in steppe and meadow areas (Chenopodiaceae, *Sparganium*, *Artemisia*, Compositae, and others; Panova, 1979).

TAPHONOMY

The common feature of the formation of early Miocene burials in the northern Aral area is the accumulation of bone remains in numerous lakes and lagoons on the coastal plain north of the North Usturt sea bay that filled the depression of the same name. The coeval existence of water basins, varying in size, depth, and salinity and having different connections with the sea and river basins, accounts for the specific diversity of facies in the Aral deposits often noted by the researchers (Yanshin, 1953; Lavrov, 1979).

The taphonomical data for the Altynshokysu locality indicate that the depositional settings of the Aral Formation became progressively more continental. As the sea retreated, all of the lagoons first became a brackish-water lake, and then gradually turned into a freshwater basin. This was responsible for the increasingly shallow-water aspect of sediment, which successively changed from silty to predominantly clayey types. The growth of the carbonate content caused the formation of intercalations of limstones and marls. The study on the taphonomic peculiarities of every bone-bearing level shows how the character of the sedimentary basin changes in the course of time (Fig. 2). As regards the actual process of the formation of burials, it appears that it did not essentially change with time, and included the transportation of bone material by water flows (temporary or permanent) and its deposition in coastal areas of water basins.

TAXONOMY

Order Insectivora Bowdich, 1821

Family Erinaceidae Fisher von Waldheim, 1817

Subfamily Erinaceinae Fisher von Waldheim, 1817

Tribe Brachyericini Butler, 1948

Genus *Exallerix* McKenna et Holton, 1967

Species *Exallerix efialtes* Lopatin, sp. nov.

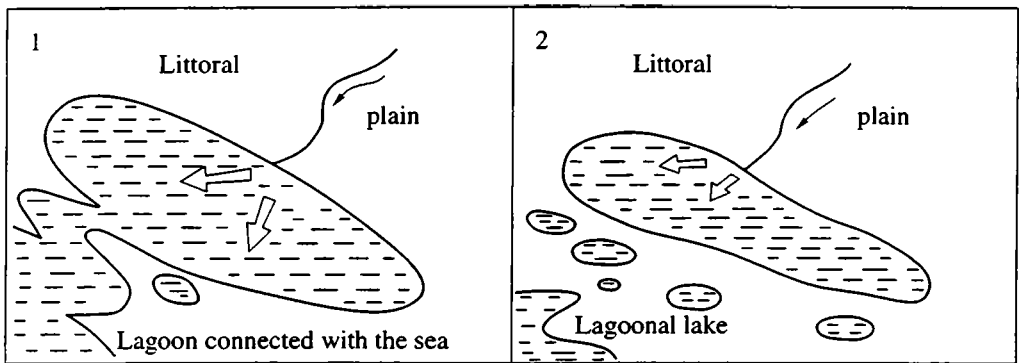
Plate, no. 6; Fig. 3

The name of the species is derived from "efialtes," which in Greek means "nightmare," or "the demon causing nightmares."

Holotype—PIN RAN, no. 4516–1; a fragment of the right bone of the lower jaw with P_4–M_2; Kazakhstan, the Northern Aral region, Altynshokysu locality; lower Miocene.

Diagnosis. The dimensions are average (P_4–M_2 is 10.5 mm long) and comparable with those of *E. hsandagolensis* McKenna et Holton, 1967. P_4 is greatly reduced (2.75 times shorter than M_1). M_1 is fairly large and narrow, with a cutting paralophid and a rather small metaconid. M_2 is small with a flattened paraconid. The size of the talonid is a little less than a half of the total length of the tooth. The masseteric ridge of the lower jaw is very massive and broad.

Description. The lower jaw has a very short dental arm, and the coronoid process is at right angles to the arm. The masseteric fossa is very deep; the masseteric ridge is powerfully developed, and its anterior border is below P_4. The mental foramen, large and funnel-shaped, is located under the anterior part of M_1. The alveole I_1 is very long, and reaches the level of the posterior border of M_1. The dental row retains the root of P_3 with a pulp cavity and absolutely whole slightly abraded P_4, M_1, and M_2. The crown of the double-rooted P_4 is strongly reduced (the paraconid and the



Stage	Bone-bearing level	Bed
1.	I, II	1, 2
2.	III	6
3.	IV	8

Fig. 2. The scheme of development of the lacustrine–lagoonal basin in the Altynshokysu area in Aralian time: a—direction of remain drifting with water flows; b—direction of material drifting within water basins.

metaconid are absent). The height of P_4 is somewhat greater than that of the M_1 paraconid, which is large and has a cutting bladelike paralophid. The metaconid is reduced; it is smaller and lower than the protoconid. Its apex is slightly more advanced than that of the protoconid, and they are divided by a shallow trough. The talonid is short but high, with a well-developed hypoconid and entoconid. M_2 is reduced to almost half of M_1. The trigonid is somewhat longer than the talonid. The paraconid is flattened. The protoconid is the most massive of the cusps. The metaconid is slightly smaller than the protoconid, but almost as high. The hypoconid and entoconid are relatively small and low. P_4 and both molar teeth have labial cingulum with a distinct relief of small rounded cusps arranged lengthwise in a single row (Fig. 3).

Table 2 shows the size of the lower jaw and teeth.

Comparison. Exallerix hsandagolensis McKenna et Holton, 1967 differs from the type species by a relatively smaller P_4 and a larger M_1 (P_4 of *E. hsandagolensis* is only twice as short as M_1), by the cutting paralophid and greatly reduced metaconid of M_1, and by the

Table 2. The size of the lower jaw and teeth of *Exallerix efialtes* sp. nov. (in mm)

Tooth	Length			Width	Height				
	total	trigonid	talonid		Pcd	Prcd	Mcd	Hcd	Ecd
P_4	2.0	–	–	2.0	–	2.5	–	–	–
M_1	5.5	4.0	1.5	2.6	2.25	3.75	2.25	2.3	1.7
M_2	3.0	1.75	1.25	2.25	1.0	2.25	1.5	1.5	1.5

Length P_4–M_2 is 10.5

Distance from M_2 to the anterior border of the coronoid process is 1.75

Height of the dental branch on the labial side: under M_1 is 6.1

under M_2 is 7.5

Note: Pcd–paraconid, Prcd–protoconid, Mcd–metaconid, Hcd–hypoconid, Ecd–entoconid.

much more developed masseteric ridge of the lower jaw.

Discussion. McKenna and Holton (1967) refer the genus *Exallerix,* defined by them, to a special subfamily of short-muzzle hedgehogs Brachyericinae, including *Brachyerix* Matthew, 1933, *Metechinus* Matthew, 1929, and *Dimylechinus* Hurzeler, 1944. Gureev (1979) attributes this group to the rank of tribe Brachyericini in the subfamily Erinaceinae. All these genera are characterized by the reduction of teeth rows: the 1–2 premolars in the lower jaw and 2–3 in the upper jaw, M_3 is absent. The number of molars of *Exallerix* is still uncertain, because it was defined in a single fragment of the lower jaw of a type species (the anterior part of the teeth arm of the holotype is broken off right behind the paraconid M_2). Consequently, some of the researchers consider the taxonomic position of *Exallerix* to be uncertain; e.g., Carroll (1988) defines it as Erinaceoidea incertae sedis. The teeth row of *E. efialtes* sp. nov. is completely preserved. The presence of two molars confirms the correctness of the previously established teeth formula for *Exallerix* ($I_2C_1P_2M_2$) and refers it to Brachyericini. Moreover, the PIN RAN collection has several fragments of the lower jaws of *E. hsandagolensis* from the middle Oligocene of Mongolia (Shand Gol Formation; collections of the Joint Soviet–Mongolian Paleontological Expedition, JSMPE), which show that this species does not have M_3.

The presence of *E. efialtes* sp. nov. in the Aral Formation extends the stratigraphic range of the genus from the middle Oligocene to lower Miocene. As compared to *E. hsandagolensis*, the new species has certain progressive features, such as the large and narrow M_1 with a cutting paralophid, the reduced metaconid and P_4, and a powerful masseteric ridge, which, as I believe, indicates the adaptation to mainly carnivorous habit.

Distribution. The species is recovered only from the Altynshokysu locality, the Northern Aral region, Kazakhstan; lower Miocene.

Material. Only a holotype is available. The jaw and teeth are well preserved.

Order Rodentia Bowdich, 1821

Family Cricetidae Rochebrune, 1883

Subfamily Eucricetodontinae Mein et Freudenthal, 1972

Genus *Eumyarion* Thaler, 1966

Species *Eumyarion tremulus* Lopatin, sp. nov.

Plate, nos. 1–3; Fig. 4a–4f

Eucricetodon aff. *youngi* Li et Qiu, 1980—Bendukidze, 1993.

The name of the species is derived from the Latin "tremulus" (shivering, trembling).

Holotype—PIN RAN, no. 4516-70; the right M_1; Kazakhstan, the Northern Aral region, Altynshokysu locality; lower Miocene.

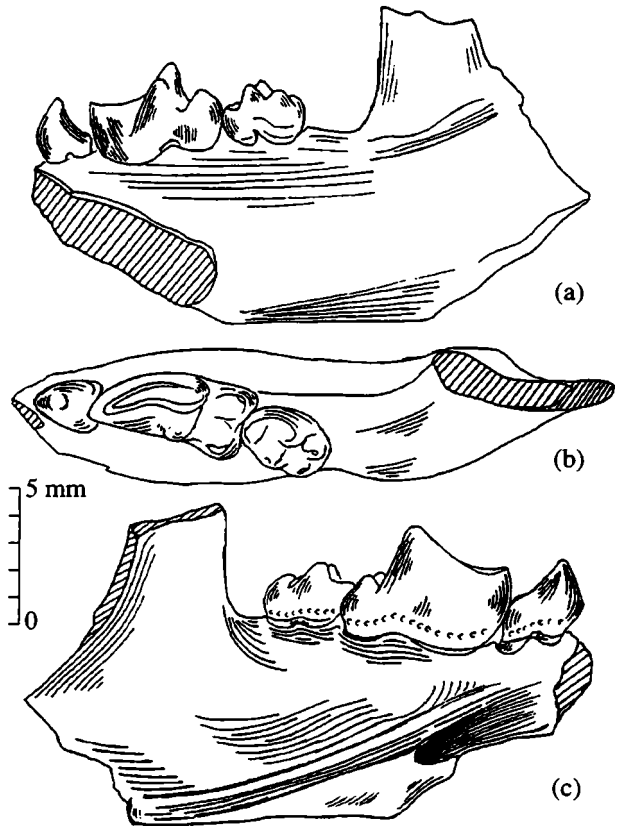


Fig. 3. *Exallerix efiaters* sp. nov. no. 4516-1, holotype, the right fragment of the lower jaw: (a) lingual side; (b) occlusal side; (c) labial side.

Diagnosis. The size is average for the genus. M_1 is 1.5–1.85 mm long of elongated–triangular shape, with isolated anteroconid and well-developed mesolophid and ectomesolophid. The posterior branch of M_1 hypoconid is rudimental; it is absent on M_2 and M_3. The anteroconid of M^1 is slightly forked; there are also anterophule, the anterior arm of protocon, the labial calcar of anterocon, the posterior calcar of paracon, and the short mesoloph. M^2 has a paracon calcar and a relatively long mesoloph. The mesoloph of M^3 and the posterior arm of the M_3 protoconid reach the border of the occlusal surface of the corona.

Description. The shape of M^1 is pearlike, asymmetric, and angular. All cusps are well developed. The anterocon is transversally elongated and occupies the central position to the longitudinal axis of the tooth, or is somewhat displaced toward the external cusps. The protosinus and anterosinus are divided by anterolophule, which joins the anterior process of the protocon. On the outside of the latter, there is a distinct calcar of the anterocon; the calcar sharply bends labially, sometimes reaching the labial cingulum. The calcar can also be straight or double-branched, and often joins the anterior process of the protocon. The paracon almost invariably has a posterior calcar. The metalophule is

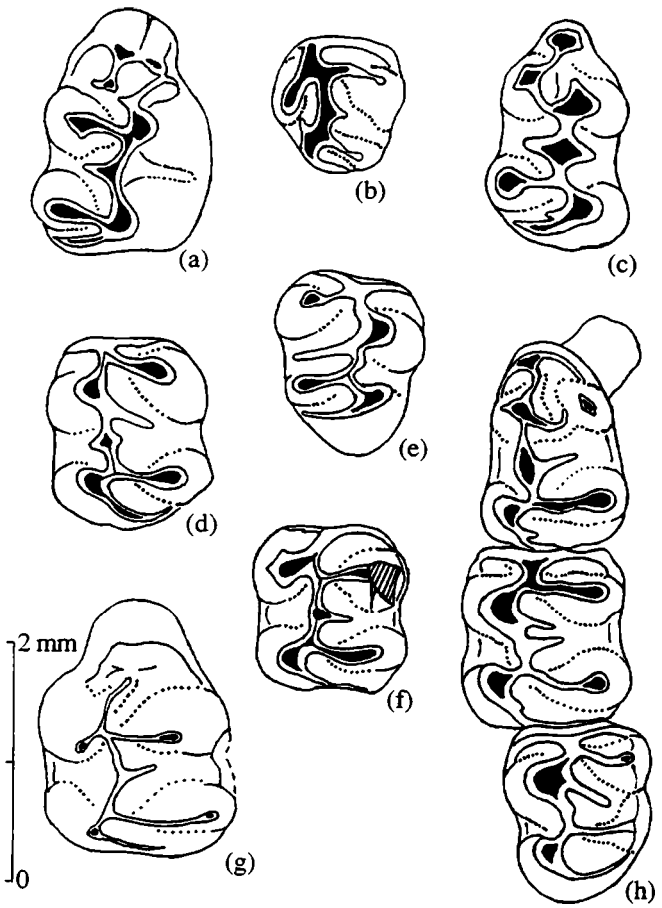


Fig. 4. The teeth of representatives of Eucricetodontinae from the Aral Formation at Altynshokysu: (a–f) *Eumyarion tremulus* sp. nov.: (a) no. 4516-60, right M^1; (b) no. 4516-389, left M^3; (c) no. 4516-70, holotype, right M_1; (d) no. 4516-15, left M_2; (e) no. 4516-402, right M_3; (f) no. 4516-46, left M^2; (g–h) *Eucricetodon occasionalis* sp. nov.: (g) no. 4516-198, left M^1; (h) no. 4516-196, holotype, left M_1–M_3.

formed by the arm of metacon joining the anterior arm of hypocon. From the place where the protocon and paracon branches merge, an oblique or almost straight longitudinal crest (entoloph) projects and joins the hypocon bearing a well-developed short mesoloph, which is parallel to transversal crests. The posterior cingulum is separated from the metacon by a narrow deep posterosinus.

M^2 is rounded–square in shape. The external branch of the anterior cingulum is well developed; the internal branch is almost reduced. The cusps are massive; the transversal crests are almost straight. The paracon has a distinct posterior calcar. The mesoloph is narrow, tapering, usually long, and reaches the calcar of the paracon. The posterior cingulum is well developed. The posterosinus is deep, long, and rather wide.

M^3 is very small, rounded, with well-developed anterior elements of an occlusal surface (the external branch of the anterior cingulum, protocon, paracon, and mesoloph), and the reduced metacon and hypocon. The mesoloph, as a rule, reaches the labial border of the occlusal surface.

M_1 is elongated–triangular with a broadening posterior heel. The anteroconid is small, well separated, and joins the metaconid. The anterior process of the protoconid reaches the base of the anteroconid; the posterior branches of the protoconid and metaconid form the slightly backward bending metalophulid. The hypolophulid is formed by the branches of the hypoconid and entoconid directed slightly forward. The oblique ectolophid is directed from the point where the protoconid merges with the arms of the metaconid, toward the convergence point of the hypoconid and entoconid arms; it has a well-developed mesolophid and ectomesolophid turned toward the anterointernal and posteroexternal sides of the tooth. The posterior cingulum is long, archlike, and separated from the entoconid by a rather vast posterosinusid occasionally crossed by a rudimentary crest of the posterior branch of the hypoconid.

The rectangular M_2 has a well-developed anterior cingulum with two branches. The straight metalophulid is formed by the anterior branches of the protoconid and metaconid. The posterior arm of the protoconid is long, and situated at right angles to the longitudinal axis of the tooth. The hypolophulid is straight. The short oblique ectolophid occasionally carries the poorly developed mesolophid and (or) ectomesolophid. The posterior cingulum is long, straight, and separated from the entoconid by a deep posterosinusid.

The angular M_3 is elongated–triangular; it narrows in the posterior part. The anterior cingulum has two developed branches. The crests join it with the metaconid and protoconid. The free posterior branch of the protoconid is long and, as a rule, reaches the border of the occlusal surface of the tooth. A rudimentary mesolophid is seldom present (Fig. 4).

The dimensions are shown in Table 3.

Comparison. The new species differs from *E. latior* (Schaub et Zapfe, 1953) by the presence of a massive posterior calcar of paracon of the upper molars, by the slightly forked anterocon of M^1, and by the covered lingual sinus of M^3. It also differs from *E. bifidus* (Fahlbush, 1964) by the more delicate anterocon of M^1, and from *E. medius* (Lartet, 1851) and *E. leemanni* (Hartenberger, 1966) by small low-corona teeth. The new species is larger than *E. microps* Bruijn et Sarac, 1991, and *E. intercentralis* Bruijn et Sarac, 1991; unlike these species, it lacks the developed posterior branch of hypoconids of the lower molars. The species differs from *E. montanus* Bruijn et Sarac, 1991, by a massive posterior calcar of paracon of the upper molars, and from *E. carbonicus* Bruijn et Sarac, 1991, by the absence of the posterior branch of the hypoconid of M_2 and M_3, by its rudimentary state on M_1, by the longer posterior arm of the protoconid of M_3, and by the less developed mesolophid and ectomesolophid of M_2.

Discussion. The *E. tremulus* sp. nov. is the first authentic find of the representatives of the genus *Eumyarion* in the lower Miocene of Asia (except Turkey). Having, in general, the primitive teeth structure similar to that of *E. carbonicus* from the Harami I locality in Turkey (early Miocene, MN 1 or 2), the species has completely lost the posterior branch of the hypoconid on M_2 and M_3, whereas on M_1 there is sometimes an almost reduced crest, which is typical of the middle Eocene *Eumyarion*. This fact apparently indicates the independent development of this *Eumyarion* lineage, and confirms the supposed Asiatic origin of the genus (Bruijn and Sarac, 1991).

Distribution. Kazakhstan, the Northern Aral area; lower Miocene.

Material. Besides the holotype, 136 teeth and their fragments were recovered from the lower bone-bearing level of the Altynshokysu locality; 36 of them are M^1, 14 are M^2, 7 are M^3, 31 are M_1, 30 are M_2, 18 are M_3; one M^1 is from the Akespe locality.

Genus *Eucricetodon* Thaler, 1966

Species *Eucricetodon occasionalis* Lopatin, sp. nov.

Plate, nos. 4, 5; Fig. 4g, 4h

Eucricetodon aff. *caducus* (Shevyreva, 1967)—Bendukidze, 1993.

The name of the species is derived from the Latin "occasionalis" (incidental).

Holotype—PIN RAN, no. 4516-196; a fragment of the left bone of the lower jaw with M_1–M_3; Kazakhstan, the Northern Aral region, Altynshokysu locality; lower Miocene.

Diagnosis. The small representatives of the genus (M_1–M_3 are 4.9 mm long; M_1 is 1.5–1.75 mm long) have rather low and broad coronas of the molars. M_1 is rounded–triangular, and the anteroconids are poorly isolated. On M_1 and M_2, the posterior arm of the protoconid, the mesolophid, and the posterior arm of the hypoconid are well developed. The posterior arm of the M_3 protoconid is short. The anterocon of M^1 is low, not forked. The mesoloph of M^1–M^3 is short. All folds are shallow, but rather long.

Description. M^1 is rounded-triangular with a central anterocon, which is small and slightly detached from the rest of the cusps; it is small and very low, with a single apex. The anterior part of the protocon is free and does not reach the base of the anterocon. The transversal crests are almost straight. The entoloph is slightly inclined, directed from the junction of the arms of the protocon and paracon toward the hypocon. The mesoloph is straight and short. The posterior cingulum is narrow and underdeveloped. The posterosinus is narrow and long.

M^2 is rounded–square. The external arm of the anterior cingulum is well developed, but its internal arm is somewhat reduced. The longitudinal and transversal crests are almost straight. The mesoloph is broad, flat, short and parallel to the transversal crests. The posterior cingulum is underdeveloped. The posterosinus is long and narrow.

M^3 is small, rounded, with greatly reduced hypocon and metacon. The mesoloph is short, and does not reach the labial border of the occlusal surface.

M_1 is rounded-triangular, and broadens in the rear. The anteroconid is low, reduced, and joins the anterior arm of the protoconid. The metalophulid is U-shaped and the hypolophulid is straight. The ectolophid is straight, and stretches from the protoconid to the merging point of the hypoconid and entoconid branches. It is slightly displaced toward the external side of the tooth and has a well-developed mesolophid pointing toward the anterointernal wall. The hypoconid and entoconid are widely spaced. There is a fairly developed posterior arm of the hypoconid. The posterior cingulum is poorly developed. The posterosinusid is small, but broad and long.

M_2 is large and rectangular. Both branches of the anterior cingulum are developed. The transversal crests are slightly bent, and the ectolophid is somewhat inclined with an obvious short mesolophid. The posteroir branch of the protoconid is free and no longer than the mesolophid. The posterior cingulum is underdeveloped. Occasionally, there is a posterior branch of a hypoconid.

Table 3. The size of teeth of *Eumyarion tremulus* sp. nov. (in mm)

Tooth	Number of measurements	Length			Width		
		min	aver.	max	min	aver.	max
M^1	29	1.75	1.9	2.05	1.1	1.25	1.35
M^2	13	1.35	1.45	1.55	1.2	1.3	1.4
M^3	6	1.0	1.1	1.2	1.05	1.15	1.25
M_1	18	1.5	1.7	1.85	1.0	1.1	1.15
M_2	20	1.4	1.5	1.65	1.05	1.2	1.3
M_3	13	1.4	1.45	1.5	1.1	1.15	1.2

Table 4. The size of teeth and of the lower jaw of *Eucricetodon occasionalis* sp. nov. (in mm)

Tooth	Number of measurements	Length			Width		
		min	aver.	max	min	aver.	max
M^1	4	1.7	1.8	2.05	1.1	1.15	1.5
M^2	2	–	1.45	–	1.25	–	1.3
M^3	1	–	1.0	–	–	1.0	–
M_1	3	1.5	1.65	1.75	1.0	1.15	1.2
M_2	2	1.4	–	1.6	1.15	–	1.35
M_3	2	1.25	–	1.4	1.0	–	1.25

Note: The length of M_1–M_3 is 4.9; the height of the dental branch is 3.8 under M_1, 3.9 under M_2, and 4.0 under M_3.

M_3 is rounded-triangular, short, slightly narrowing at the rear. Both branches of the anterior cingulum are developed. The protoconid and metaconid are widely spaced; the posterior branch of the protoconid does not reach the wall of the tooth by a quarter of the corona width. The hypoconid and the posterior cingulum are well developed; the entoconid is considerably reduced.

The dimensions are given in Table 4.

Comparison. The small uniapexed anterocon distinguishes the new species from *E. gerandianus* (Gervais, 1848), *E. infralactorensis* (Viret, 1930), *E. aquitanicus* Baudelot et Bonis, 1968, *E. occitanicus* Bonis, 1970, *E. hurzeleri* Vianey-Liaud, 1972, *E. cetinensis* Daams, 1976, and *E. margaritae* Daams *et al.*, 1989. The short mesolophs of M^1–M^2, the absence of the mesolophid of M_3, and the weak posterior cinguli of the new species distinguish it from *E. collatus* (Schaub, 1925), *E. dubius* (Schaub, 1925), and *E. longidens* Hugueney, 1969. The absence of the M_1 ectomesolophid and M_3 mesolophid distinguish *E. occasionalis* sp. nov. from *E. atavus* (Misonne, 1957), *E. praecursor* (Schaub, 1925), *E. asiaticus* (Matthew et Granger, 1923), and *E. caducus* (Shevyreva, 1967). The small size and less massive cusps distinguish it from *E. robustus* Agusti et Arbiol, 1989; the greatly reduced posterior part of M_3 from *E. gergovianus* (Gervais, 1848) and *E. huberi* (Schaub, 1925); the small anterocon and short mesolophs and mesolophids from *E. incertus* (Schlosser, 1884), *E. hochheimensis* (Schaub, 1925), *E. haslachensis* (Schaub, 1925), *E. leptaleos* Wang et Meng, 1986, and *E. sajakensis* Bendukidze, 1993. A relatively larger anteroconid distinguishes the new species from *E. schaubi* (Zdansky, 1930); the absence of mesostyls distinguishes it from *E. meridionalis* Wang et Meng, 1986; the long posterosinuses, straight protolophule and more developed lingual branch of the anterior cingulum of M^2 distinguish it from *E. youngi* Li et Qiu, 1960; and the absence of calcar of the anterocon along with the larger size from *E. murinus* (Schlosser, 1884); from *E. hesperius* Engesser, 1985 it differs because of a smaller anterocon, weaker posterior cinguli, and posterior branch of the hypoconid of M_1–M_2.

Distribution. Kazakhstan, the Northern Aral area; lower Miocene.

Material. Besides the holotype, there is one fragment of the left bone of the lower jaw with M_1, an isolated M^1 recovered from the upper bone-bearing level of the Altynshokysu locality, and 12 teeth from the lower level: 5 of them are M^1, 2 are M^2, 1 is M^3, 2 are M_1, 1 is M_2, and 1 is M_3.

CONCLUSION

Two assemblages different in age and composition were defined as a result of the study on the small mammal remains recovered from the Altynshokysu locality. The first assemblage from the lower bone-bearing level Altynshokysu (I) is of the early Aquitanian age. It is correlative with the MN 1 zone of the Mein's scale and with the Akespe fauna. The mammal assemblage from the upper bone-bearing level Altynshokysu (IV) can be compared with the fauna of other localites on the northwestern coast of the Aral Sea. The age of this complex is the late Aquitanian.

The described mammalian remains from the Altynshokysu locality indicate the considerable diversity of the Aral fauna and its similarity with the late Oligocene and early Miocene assemblages of China. Three new taxa were distinguished: *Eumyarion tremulus* sp. nov., *Eucricetodon occasionalis* sp. nov., and *Exallerix efialtes* sp. nov. The peculiar features of many other small mammals suggest that further research will show how these species differ from the cognate Oligocene forms. These results would be useful for a more fundamental understanding of the development history of the Aral fauna.

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Reviewer M.A. Akhmet'ev.

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Problems of the Caucasus Paleoland Existence during the Oligocene–Miocene Time

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Abstract—This paper considers the litho-facial environments that existed in the Caucasus region during the Oligocene and Miocene. The analyzed distribution patterns of reef deposits, evaporites, rock slide and avalanche facies, coal deposits, and paleoflora provided, in addition to denudation chronology, the basis to argue that the Caucasus orogen of that time existed as an island, actively influencing the sedimentation in adjacent basins.

Key words: *lithology, paleogeography, Maikop Group, Middle Miocene, Caucasus Island.*

The problem of whether or not the Caucasus landmass existed during the Tertiary time was always topical in paleogeography, geology, eventual stratigraphy and tectonics. One solution or another of the problem was of crucial importance for positioning boundaries between stratigraphic subdivisions, outlining limits of zoo- and phytogeographic provinces, and understanding the evolutionary patterns of folding events and the Caucasus orogenic system in general.

For a long time, it was accepted that the Caucasus Island came to exist during the Oligocene. Zhizhchenko (1940), Khain and Leont'ev (1947, 1950), Vassoevich (1952, 1959), Vassoevich and Grossgeim (1951), Vinogradov *et al.* (1968), Strakhov *et al.* (1968), Kholodov and Nedumov (1981), and Khain (1994a, 1994b) always marked an island area, the protoland of the central Caucasus, in paleogeographic maps of the Oligocene time. Similar areas were also outlined for the Early, Middle, and Late Miocene substages reflecting the persistent upraise of this region.

Nevertheless, this seemingly valid idea was recently revised. Onishchenko (1974), followed by Shcherba (1987), argued that the Caucasus landmass hardly existed during the Oligocene, when, according to their paleogeographic reconstructions for this and later intervals, the land areas were scattered only along the marginal part of the Russian platform and in Transcaucasia.

We believe that such a categorical conclusion is insufficiently verified by litho-facial and paleogeographic data, and despite the attractive novelty, it should be repudiated.

In the present analysis of factual materials elucidating the history of Caucasus Island, it would be most convenient to begin with data concerning the Middle Miocene, which corresponds to the formation epoch of the Chokrak and Karagan deposits.

We investigated the Middle Miocene deposits in eastern Ciscaucasia, precisely in the Urukh–Sulak interstream area (Kholodov and Nedumov, 1981). Studying outcrops along the Urukh, Suadag-Don, Kodakhchin, Buivolinaya, Fortanga, Malaya Roshnya, Elistanzhi, Yaryk-Su, and Sulak rivers, we described and intercorrelated nine complete sections (Fig. 1). The "conglomerate β," well traceable all along the profile (Vassoevich, 1952) was used as the marker horizon.

The most representative Middle Miocene section is exposed along the Sulak River. Here, the Chokrak deposits consist of alternating, brownish gray and black, thin-bedded marine clays enclosing abundant calcareous and sulfide concretions with the following fauna: *Spirialis* cf. *subtarchanensis* Zhizh., *Cryptodon sinuosis* (Don.) Andrus., *Arba alba scythica* (Sok.), *Nassa restitutiana* Font., *Buccinum dujardini* Desh., and *Leda prendeli* Andrus. In addition, the deposits include marl and thick sandstone interbeds (beds Σ_1–Σ_7), which yielded the following molluscs: *Donax tarchanensis* Andrus., *Lucina* sp., *Dosinia* sp., *Venus* sp., *Solen* sp., *Ervilia passilla praepodolica* Andrus., *Mactra bajarunasi* Koles., *Chlamys*, and some others. The deposits are 640 m thick.

The Karagan deposits also consist of alternating black and brownish gray clays with abundant interbeds of marl and white oligomictic quartz sandstone (S_1–S_{13}). These rocks are completely barren of foraminifers and *Spirialis* forms, though they do contain molluscs represented by specimens of *Spaniodontella, Sandbergeria, Mohrensternia*, and *Barnea* genera, with the first of them as the most abundant. The total thickness of these deposits is 350 m.

On going from the eastern localities westward, we observed in the Middle Miocene sections some facial changes typical of the transition from deep- to shallow-water settings. Westward of the Fortanga River, their

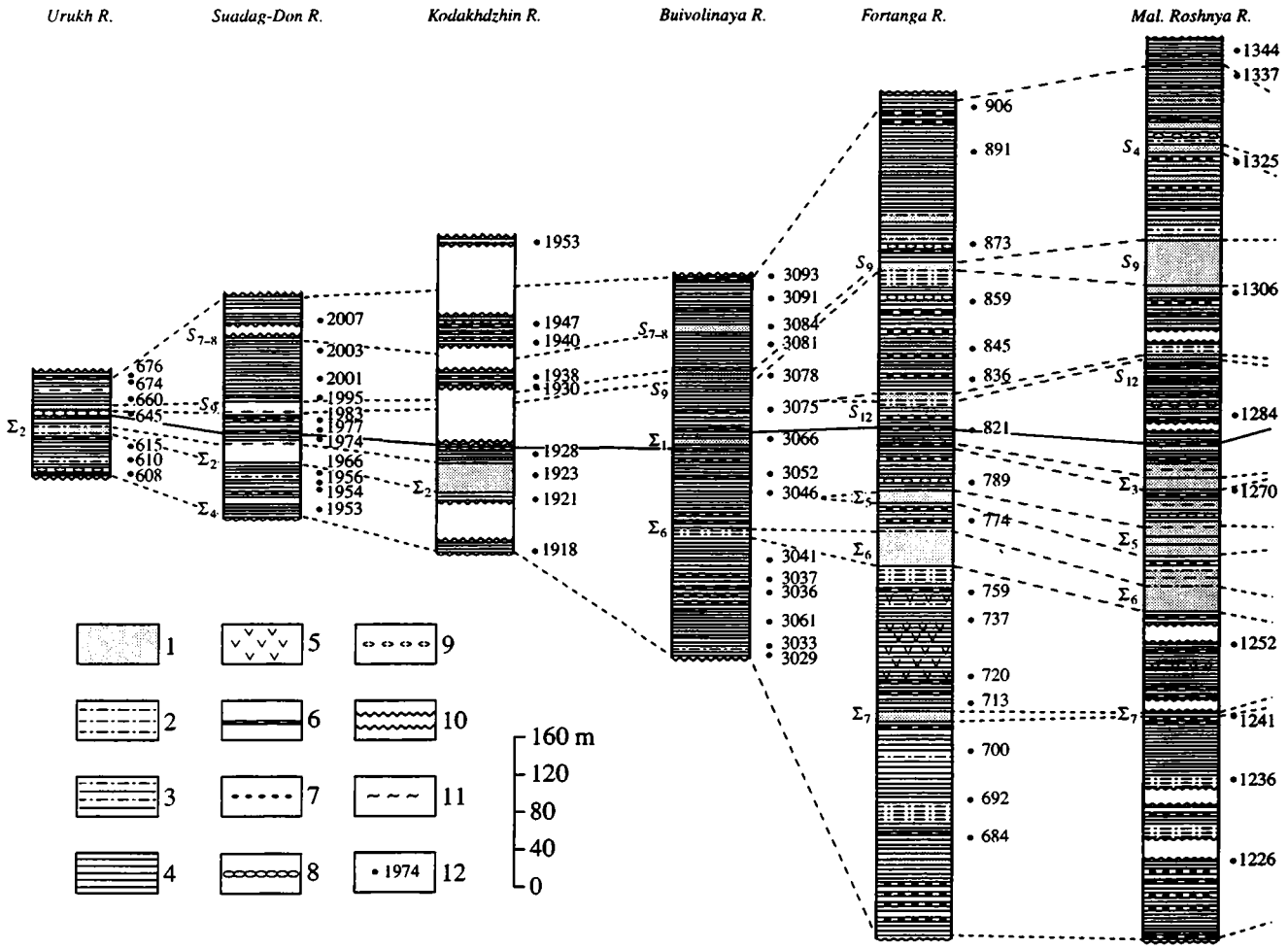

Fig. 1. Facial changes and stratigraphy of the Middle Miocene deposits in eastern Ciscaucasia.
1—sandstones; 2—siltstones; 3—clayey siltstones; 4—clays; 5—gypsum-bearing clays; 6—marls; 7—calcareous concretions; 8—stromatolites; 9—coquinas; 10—unexposed intervals; 11—soft sediment deformations and textures of submarine landslides; 12—point of sampling. Symbols Σ and *S* mark sandstone beds.

thickness decreases down to 400 m; the abundance of clays increases here, and oligomictic sand and sandstone beds disappear from the sections.

In the westernmost section of the Middle Miocene deposits along the Urukh River, the summary thickness of Chokrak and Karagan deposits decreases down to 120 m. The basal part of the Chokrak regiostage here yielded fresh-water fauna (*Planorbis, Viviparus, Pupa, Hydrobia*, and some other forms) occurring in beds of green chlorite clay. The unit locally includes coarse-grained sands and conglomerates, and the Chokrak–Karagan boundary is marked by the well traceable horizon of stromatolite limestones.

This stromatolite horizon is clearly distinguishable in an exposed section of the Keta-Cheta Gully. In this locality, a conglobreccia bed of variable thickness occurs 10–11 m above the bed of white quartz sandstone (Σ_1 or Σ_2?), where the former rests on the layered brownish gray clays of the Chokrak regiostage (Fig. 2). The conglobreccia consists mostly of pebbles and fragments of organogenic algal limestone; flat pebbles of marl occur less frequently. The pebble size is usually 3–4 cm, though occasional pebbles are as large as 10 cm in diameter. The clayey rock matrix is highly ferruginous and brown in color. The thickness range of this locally pinching conglobreccia bed is normally between 20 and 80–90 cm.

Large organogenic algal nodules and bioherms (size range from 0.1 × 0.3 to 0.5 × 0.8 m) are confined to the uppermost part of the bed. These bodies, sometimes isolated, usually form lenticular segregations up to 2–3 m in length. Some outcrops clearly show that several bioherms in these segregations are not in their original position, as they are somewhat rotated and displaced relative to one another.

Dissected stromatolite bodies demonstrate an inner core consisting of small algal nodules, pebbles, and fragments, all surrounded with a calcareous envelope. The latter is sometimes curling around the central core following the roll, or snow ball patterns (Fig. 2).

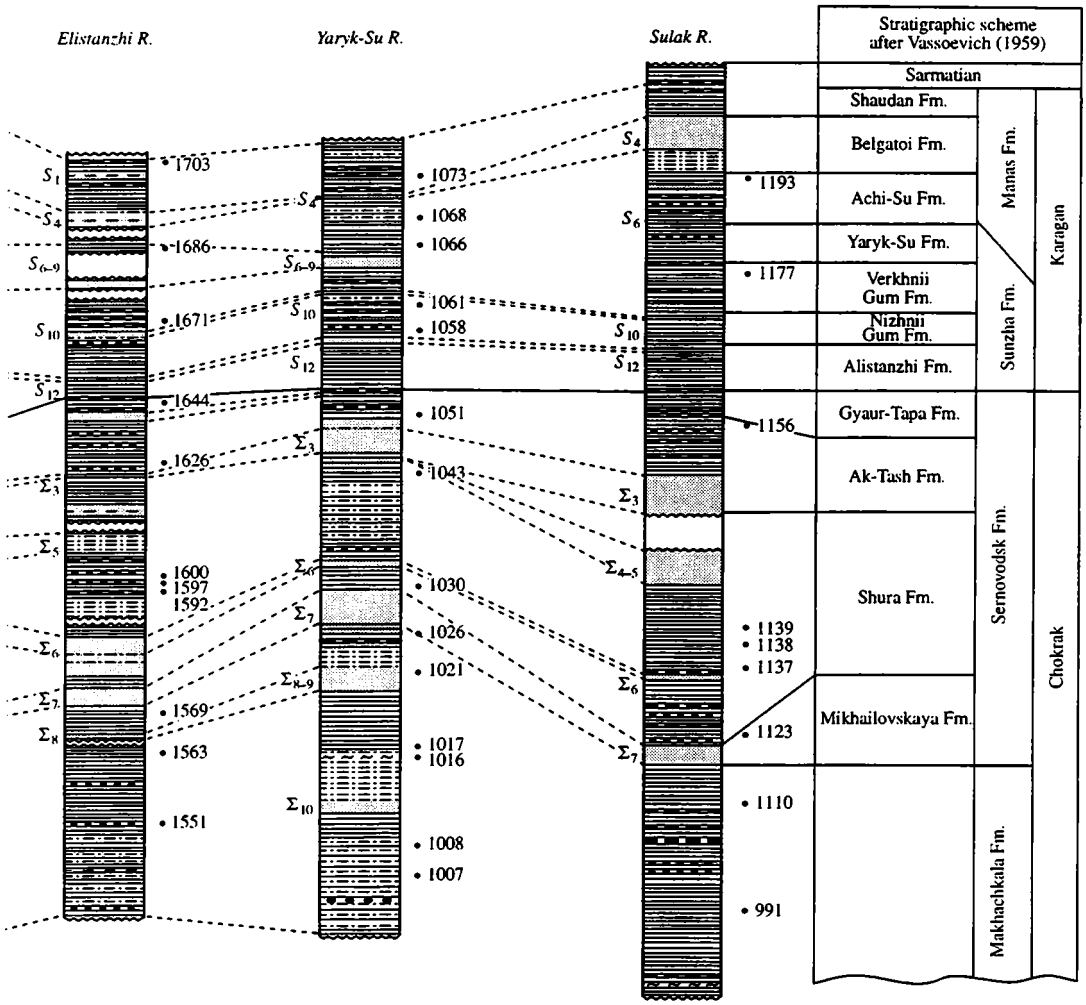

Fig. 1. (Contd.)

In the clays enclosing the organogenic algal bodies, bedding being disturbed nearby partially coat them, as it is illustrated in Fig. 2. The outer surface of bioherms and nodules is either smooth, or tubercular; sometimes it resembles that of cerebrum-like forms and usually is strongly ferruginated presumably because of pyrite oxidation.

We also traced the stromatolite horizon along the right side of the Urukh River near the Arkharisar, and in sections of the Suadag-Don River. Beyond a doubt, this horizon represents the most shallow-water facies of the Middle Miocene basin corresponding to the bryozoan-algal limestones of western Ciscaucasia.

The high abundance of green chlorite clays is another typical feature of the Urukh section. An x-ray diffraction study (Kholodov and Nedumov, 1981) demonstrated that these clays consist of montmorillonite (20–30%), hydromica (40–70%), and Mg–Fe-chlorite (10–20%). To explain the chlorite appearance within the coastal zone of the studied profile, we have to take for its provenance the eroded weathering crust of the ultrabasite Malka Massif, as it was precisely in this massif that Serdyuchenko (1953) distinguished a complete series of Mg–Fe-chlorites, the alteration products of ultrabasic magmatic rocks. Therefore, we have all the necessary grounds to suggest that southwestward of the Urukh locality, there was a Middle Miocene land area, which provoked the influx of fine clay particles into the Chokrak basin.

Thus, the litho-facial analysis show that two features accurately indicate the extremely shallow-water paleosettings: the appearance of organogenic algal limestones in the sections and abundance of green chlorite clays in the rock composition.

Seeking this combination of litho-facial indicators in the Middle Miocene deposits of the western Caucasus, we easily come to the conclusion that respective shallow-water facies indicating the proximity of settings to the coastline occur nearly everywhere in this region.

Actually, the bryozoan reef buildups and *Chama* limestones (Prokopov, 1914; Zhizhchenko, 1940) are the widespread constituents of the Chokrak deposits in

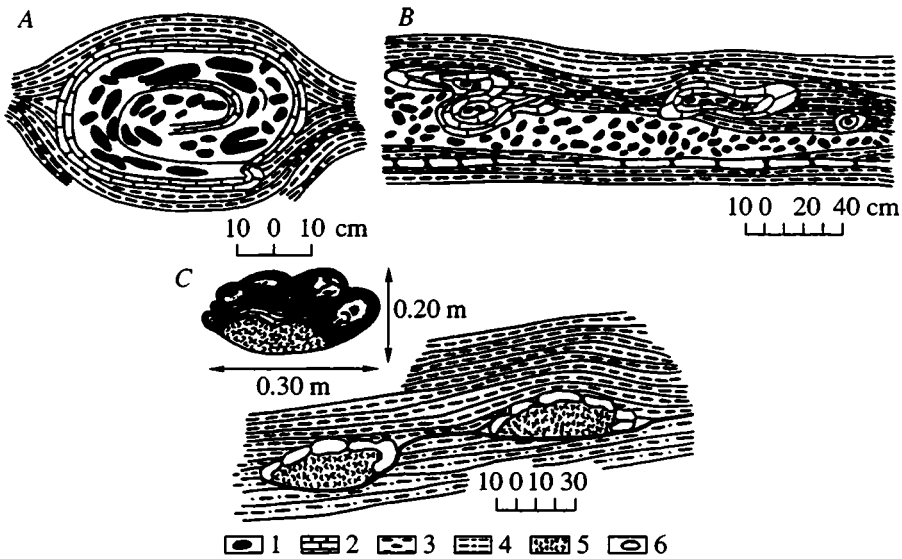

Fig. 2. Types of the Middle Miocene stromatolite buildups occurring in eastern Ciscaucasia.
A—section of a bioherm; *B*—flat-pebble conglomerates and bioherms in the Urukh section; *C*—coquina with biohermal rims in the Khatal-Don section; 1—pebbles and fragments of algal limestones; 2—algae-serpulid rims; 3—clay; 4—siltstones; 5—coquina composed of *Spaniodontella* shells; 6—serpulid nodules.

the northwestern part of the region near the Kaluzhskaya Station (Seversk district) and city of Khadyzhensk (Apsheron district). Both of these rocks are facial analogues of organogenic algal sediments occurring in eastern Ciscaucasia. In addition, the suggested facial relations are evident from the fact that the Chokrak deposits are frequently pinched over apexes of synsedimentary anticlines distinguished by Zhabrev (1960) on the basis of prospecting drilling in the West Kuban' depression.

During the Karagan time, an area of shallow-water carbonate sedimentation was displaced northwestward, in the Anapa region. Here, near the village of Gostagaevskaya, bryozoan limestones and oolitic segregations are spatially related to lagoonal gypsum-bearing deposits (Prokopov, 1914). It is also remarkable that in some regions of western Ciscaucasia, beds separating the Chokrak and Karagan deposits contain the freshwater mollusks *Planobris* and terrestrial gastropods *Helix* (Zhizhchenko, 1940).

These data suggest that reef limestones typical of the western margin of Caucasus Island were also characteristic of its southern coast during the Middle Miocene time. For instance, in the northern Rioni depression between the villages of Indushi and Khonchiori, the Chokrak regiostage comprises bryozoan reef limestones with the abundant and diverse fauna of shallow-water mollusks (Meffert, 1930). Approximately in the same region of Mingrelia, the clay-terrigenous deposits of the Karagan regiostage are also replaced by bryozoan reef limestones, which indicate the coastline proximity (Meffert, 1931; Zhizhchenko, 1940; Vassoevich, 1958).

Isolated relics of these deposits occurring in central areas of the Greater Caucasus indicate the past limits of a land mass that existed here during the Middle Miocene. Its western relic is known in the Bol'shaya Liakhvi middle course, the Dzhava district of northern Osetiya, where it occurs in the height range of 1200–1500 m above sea level. In this locality, the Chokrak deposits are thin and consist of alternating dark gray clays and conglomerates, the latter with abundant pebbles of redeposited Mesozoic rocks. As was shown (Vassoevich, 1931, 1959), the Chokrak and Karagan deposits of similar composition were accumulated here in a local depression bounded on the north and south by past land areas.

In the east, relics of Middle Miocene deposits are known in Dagestan, in several localities of the Nukatl' Ridge, and in the interstream area of the Avarskoe Koisu and Karakoisu rivers, where (near villages of Butsra and Darada) the Chokrak deposits consist of sandstones (10–15 m in thickness) and rest on the erosion surface of the Cretaceous sediments. These shallow-water coastal deposits sporadically yielded mammalian remains, namely, the bones of Miocene goats (Vassoevich, 1959; and our finds).

In the northwestern Caucasus, the paleoland area is also marked by a distribution of saliferous and gypsum-bearing deposits. Such evaporite formations are known to be deposited under the arid climate in partially enclosed basins adjacent to continents (Strakhov, 1963; Korenevskii, 1973; Zharkov, 1978; Yanshin and Zharkov, 1986).

In the northern Caucasus (the Fortanga–Malaya Roshnya interfluve westward of the Vladikavkaz), the Chokrak deposits fill a distinct and large synsedimen-

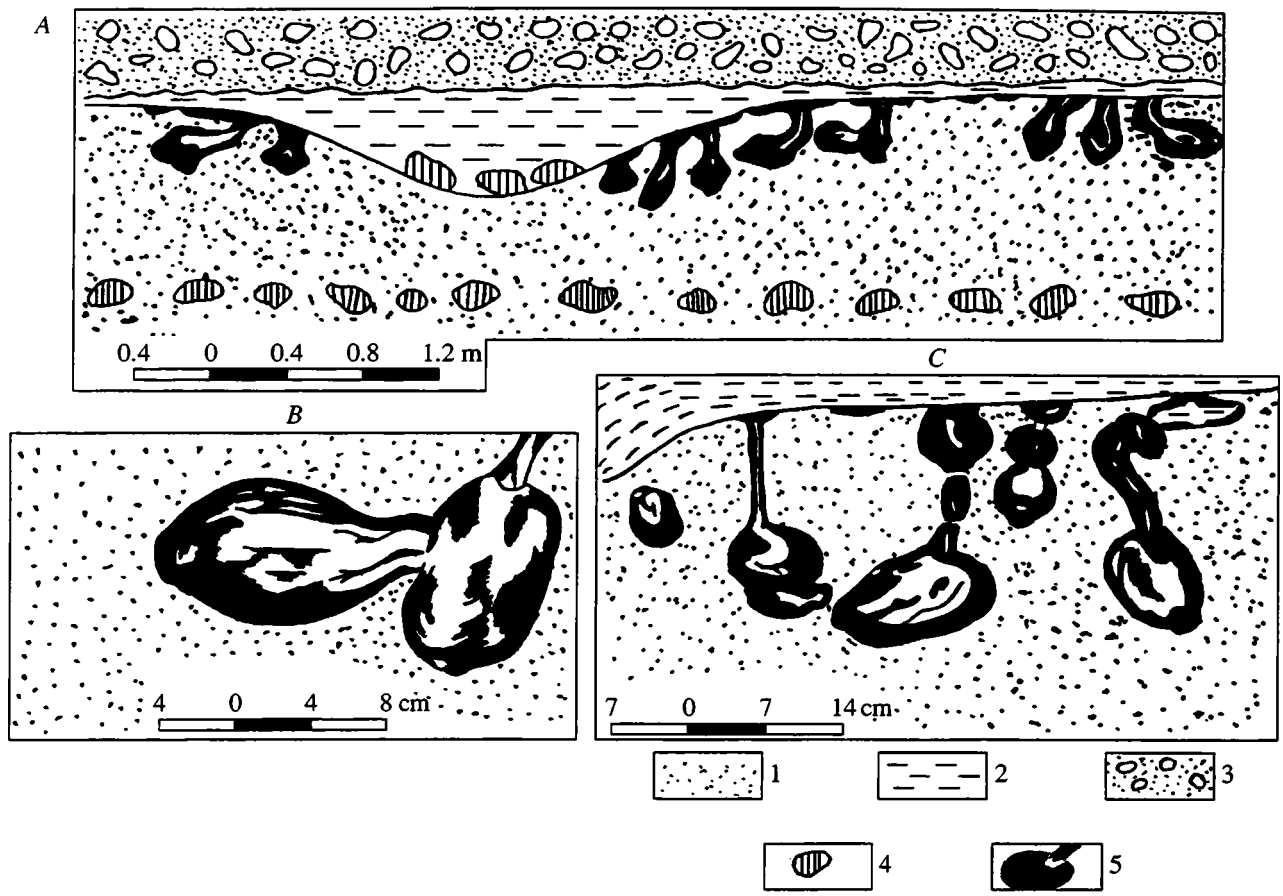

Fig. 3. The Middle Miocene tuberiferous horizon in the Fortanga section.

A—a sketch of the outcrop; *B, C*—fossil plant tubers in their natural occurrence; *D, E, F, G*—photos of plant remnants; 1—siltstones; 2—clays; 3—conglomerate of the Quaternary terrace; 4—siderite concretions; 5—tubers.

tary depression, where the Datykh Formation was accumulated. The latter is a 60-m-thick unit comprising gypsum and anhydrite beds (2–3 and sometimes 5 cm in thickness) with interlayers of gray, brownish gray, and black clays. The proportion between gypsum and clay beds varies between 1 : 1 and 1 : 2. The beds show an intense crenulation with complex folds of variable magnitudes. The Datykh Formation also contains halite preserved below the hypergenesis zone, and near the Nizhnii Datykh, there are many pits which were excavated long ago to obtain salt from natural brines (Vassoevich, 1959). The evaporite-bearing deposits extend for a distance longer than 30 km, and their outcrops are known in the Fortanga–Argun interfluve area and along the Sunzha River. These rocks were evidently accumulated in a very shallow synsedimentary basin, because they include, in addition to evaporites, separate horizons rich in fossil tubers, the remnants of aquatic plants. These remnants were discovered in siltstones of the Fortanga section below the evaporite-bearing horizons. As we already reported (Kholodov *et al.*, 1980; Kholodov and Nedumov, 1981), the tubers occur below an irregular erosion surface within the uppermost 30-cm-thick part of the siltstone bed. They are rounded, slightly elongated (Figs. 3*A*–3*D*) and variable in size (diameter up to 2–3 cm; length up to 8–10 cm). Many of them have the collar, presumably the transitional element to either the epiterranean part of plants, or the other tuberous bodies. Most of collars show subvertical orientation, thus indicating that the tuberiferous horizon represents the buried bottom area formerly overgrown with plants. The external envelope of tubers consists of organic matter; it is 0.5–0.7 mm thick. The internal part is filled with silty material showing plastic-flow textures. In some bodies, the terrigenous quartz, mica, and glauconite are cemented by pyrite. We failed to correctly identify the plants corresponding to these remains.

Among modern plants, those with tubers are typical of various genera and families of monocotyledons, especially of Orchiodaceae. For instance, the epiterranean tuber of *Coelogene cristata* Lind. is very similar to our finds. However, they could also be referred to some other families, e.g., Zingibcraceae, or Araceae.

According to Vassoevich (1959), the brownish gray clays of the Chokrak regiostage from eastern Ciscauca-

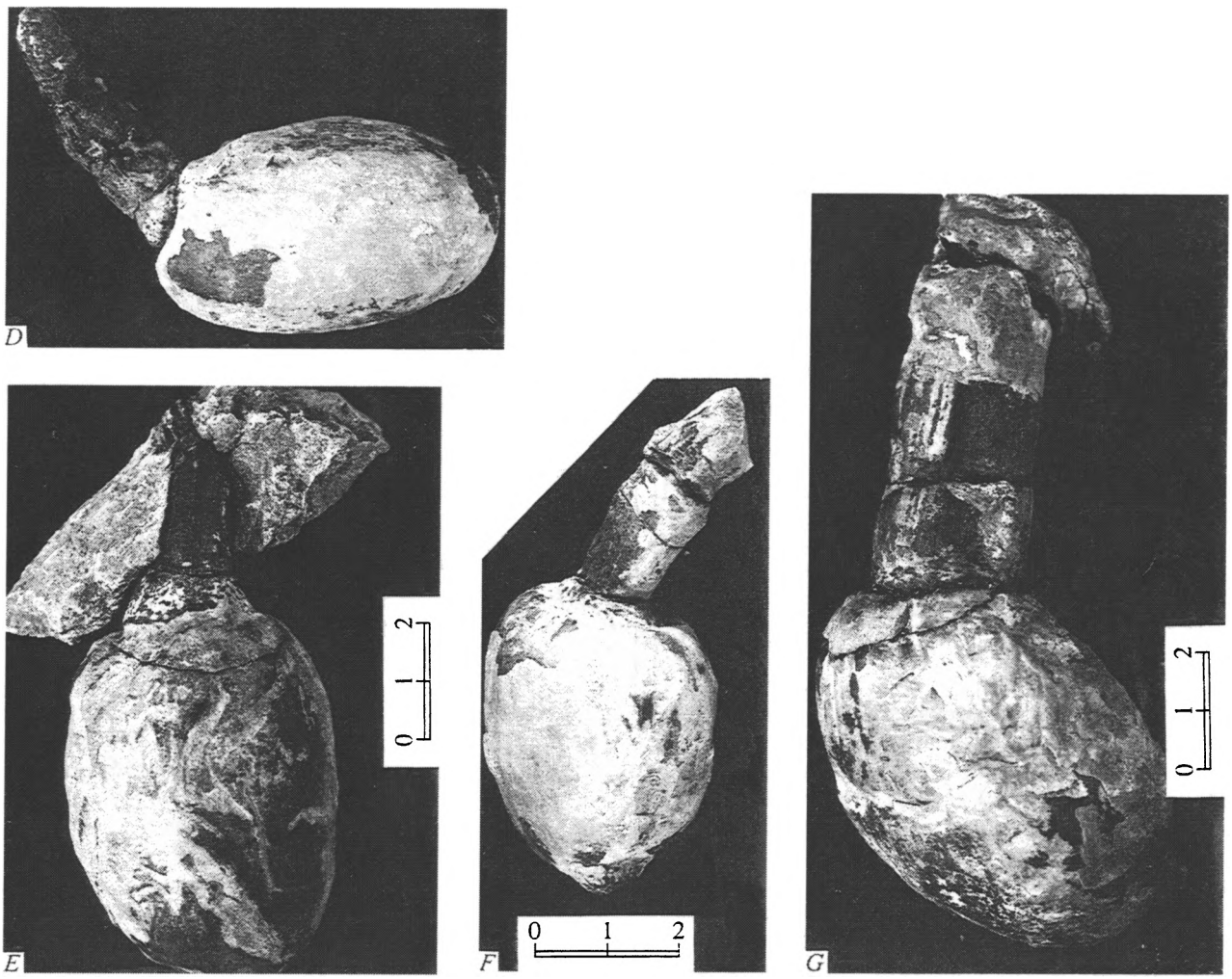

Fig. 3. (Contd).

sia may often contain the root remnants of gramineous plants (*Arundo* sp., *Typha latissima* A. Br., and other forms). They presumably were buried in abundance in shallow coastal zones of separate gulfs, or in swampy estuaries. Comparable settings correspond to present landscapes of mangrove shrubs (Nalivkin, 1956). Later, these semimarshy landscapes were transformed into deeper bays and lagoons with the evaporite sedimentation.

Evaporite horizons are also known in the Karagan deposits. As we already noted, a gypsum-bearing member over 5 m in thickness is distinguishable in the Karagan sequence of the northwestern Caucasus, near the villages of Gostagaevskaya, Krymskaya, and Varenikovskaya, where its outcrops are traceable along a distance more than 2 km (Prokopov, 1914). Similarly, lenticular interlayers of gypsum are known in the Rioni valley northward of Kutaisi, where they are up to 1 m in thickness (Strakhov, 1963).

The Middle Miocene deposits of eastern Ciscaucasia often contain coaly wood remains. Fragments of fossil trunks and branches, as well as leaf and rind imprints, are abundant in the clay and marl beds of the Chokrak regiostage, which are exposed along the Urukh, Kodakhdzin, Kambileevka, Buivolinaya, Fortanga, Roshnya, and Sulak rivers. The remains of terrestrial flora were buried in extremely shallow-water sedimentary facies which were spread around Caucasus Island. Without a doubt, this coaly material was transported from the island, rather than from the more distant platform, because the Middle Miocene succession of the Tar depression (20–25 km southeast of the Vladikavkaz along the Kambileevka River) contain deposit of brown coal. The latter is a result of the coalification of typical paralic peat soil that accumulated within a marshy paleodelta.

According to Vassoevich (1959), floral remains in the Chokrak deposits contain angiosperm forms, such as *Ulmus longifolia* Ung. and *Populus* sp. (determinations of G.N. Baikovskaya). In addition, Grossgeim and Gladkova (1950) discovered that spore–pollen assemblages from these rocks include specimens of *Juglans, Fagus, Carpinus, Betula*, and other genera

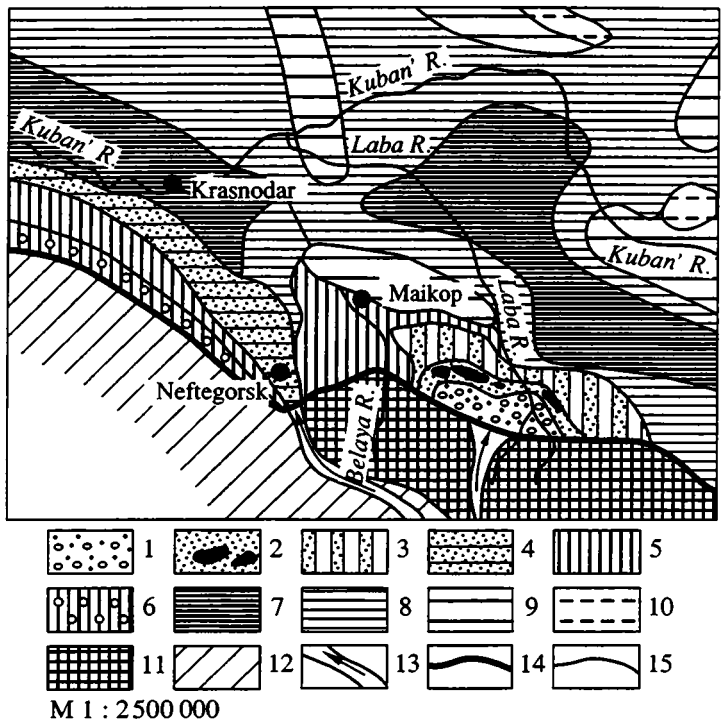

Fig. 4. Paleogeographic scheme of the Laba manganese deposit (after Strakhov *et al.*, 1968).
1—subaerial part of delta; 2—submarine part of delta; 3—avandelta; 4—distribution area of clastic material; 5—coastal clays; 6—submarine landslides; 7—abyssal clays; 8—clays of moderate depth; 9—shallow-water clays; 10—shallow-water siltstones and sandstones; 11—lowland; 12—elevated land; 13—paleoriver; 14—coastline; 15—boundary between facial zones.

typical of the Turgai floral province. Specific elements of the pollen spectrum are *Magnolia, Buxus*, and *Ilex*, the representatives of subtropical and tropical flora.

Vassoevich (1959) also reported that the clays and marls of the Karagan regiostage yielded remains of palm species (*Sabal major*), *Laurus, Juglans, Calluna*, and other forms (determinations of I.V. Palibin and G.N. Baikovskaya). Spore–pollen analysis showed that the terrestrial vegetation of that time was rich in tropical and subtropical elements, including the photophilous *Pistacia*, while coniferous (*Pinus, Picea*) and therophyllous (*Fagus, Ulmus, Juglans, Tilia*, and others) were dominant forms of the Turgai flora.

The fact that island vegetation intensively influenced sedimentation in adjacent basins of the Middle Miocene epoch is also evident from available data on fossil insects. According to Kaspiev (1939), Grossgeim and Gladkova (1950), and Vassoevich (1959), most of the fossil Arthropoda from the northern Caucasus are representatives of the mesophilous and hydrophilous insects which inhabited lagoons, bays, and marshes. There are also remarkable finds of fossil termites (*Ysoptera*), the obvious dwellers of continental arid landscapes. In her excellent work, Bekker-Migdisova (1964) described a series of homopterous insects from the Chokrak–Karagan deposits exposed near the Vishnevaya Balka, the village of Temnolesskaya, Mt. Yasenevaya, and Sangileevskoe Lake (in Stavropol'e). The discovered fossils are divisible into several past communities of plants and insects: (1) biocoenosis of vegetation typical of marine coasts and brackish-water lagoons; (2) coastal biocoenosis of fresh-water vegetation; (3) biocoenosis of inundated forests and shrubs of Turgai type; (4) biocoenosis of deciduous–indeciduous forests; (5) biocoenosis of dry and light pine forests; (6) biocoenosis of either savanna-type deciduous forests, or shrub-steppe landscapes; and (7) biocoenosis of coniferous forests.

All these assemblages of plants and insects suggest a very complex landscape structure of Caucasus Island. In our view, coastal biocoenosis characterizing lagoon, liman, and delta landscapes with reed marshes and shrubs are the most interesting. Floral remains from coastal lagoons of the Karagan sea are rare and include species of brown and red algae (*Cystoscira* sp. and *Sphaerococcites* sp.) associated with some herbaceous forms. Larvae of *Trichoptera*, some bugs (Gerridae), and dipterous forms (Chronomidae) consorting this flora indicate that the water salinity was hardly above 5‰.

Coastal plants *Najadopsis dichotoma* Heer. and *Phragmites oenirgensis* A. (gramineous) were widespread near the coastline of limans and deltas. Closer to the coast, they were associated with the fern *Dryopteris stiriaca* Ung. and sedge forms (*Carex*); further on this flora was replaced by an undergrowth of osier and alder. Mayflies (Ephemeroptera), skimmer larvae

(Odonata), water striders (Gerridae), diving and water-scavenger beetles (Dytiscidae and Hydrophilidae), caddis flies (Trichoptera), and harlequin flies and larvae (Chironomidae and Culicidae) inhabited the freshwater basins. Coastal marshes with thick gramineous vegetation (*Phragmites ceningensis* and *Carex*) were rich in various leafhoppers (*Chloriona stavropoliana* sp. and *Cicodella viridis* L.).

The shrub assemblage resembling that of present tropical marshes was dominated by *Salix lavateri* Heer. and *Alnus kefersteinii* Goepp. The respective forest landscapes were the perfect habitat for abundant psylla, soil and fungi insects, scorpion flies, butterflies and leafhoppers. In view of the above facts, the past continental uplift unquestionably existed in the Stavropol'e region.

The available data on the composition of accessory minerals from Chokrak–Karagan deposits also favor the existence of an ancient land area in the Caucasus.

According to Grossgeim and Korotkova (1957), a part of the Chokrak sea, which was between the Caucasus and Russian platform, is distinctly divisible into several litho-mineralogical provinces. Deposits of the northernmost Don province are rich in accessory tourmaline, disthene, staurolite, and black ore minerals. The southern sublatitudinal belt consists of the West Kuban', East Kuban', and eastern fore-Caucasus provinces. The main accessory mineral here is epidote associated with zircon, garnet, tourmaline, rutile, disthene, and staurolite in the west, and with disthene, staurolite, and andalusite in the east. A separating zone with a transitional composition of accessory minerals is near the village of Anastas'evskaya.

It is clear that the Russian plate was a provenance of sedimentary material for the Don province, while the West and East Kuban' provinces accumulated minerals derived from Caucasus Island. In the last case, the mineralogical difference is a result of the erosion of two different rock complexes: the Cretaceous sediments on the east and the Paleozoic formations on the west, including granites and metamorphic rocks. According to Grossgeim and Korotkova (1957), this suggests a high degree of erosion of the Caucasus landmass during the Chokrak and Karagan time. Summarizing all the available data, these researchers stressed that "... sedimentary material was transported into the Don province of the Karagan basin from the north, while a provenance for the Kuban' province was on the south."

In considering the paleogeography of the Oligocene–Early Miocene time, we have to mention in advance that many serious arguments concerning this period also favor the idea regarding the past existence of Caucasus Island.

In northwestern Ciscaucasia, precisely along the Laba River middle course, the middle–upper portion of the Maikop Group, occurring between the Batalpashinsk and Ritza formations of clayey deposits, includes a thick member of coarse deltoid sandstones and conglomerates, the host rocks for the Laba manganese deposit. The paleogeography of this area was studied in detail by Kulikov (1965), Kalinenko (1967, 1990), and Strakhov *et al.*, (1968), whose principal results are shown in Fig. 4. As one can see, near the Laba River, southeastward of Maikop, there is a junction zone between an area free of the Oligocene–Early Miocene deposits (the southern part of the scheme) and an area with coarse facies of this age (the northern sector). The latter deposits are divisible into facies of subaerial delta (sandstones and conglomerates), facies of underwater delta (sands and silts, with manganese ore bodies), and facies of avandelta grading into normal marine deposits. Characterizing them in detail, Strakhov *et al.* (1968, p. 232) reported:

> Subaerial deltoid deposits consist of fine- to medium-grained sands and sandstones associated with gravel, siltstones, gravelstones, and conglomerate beds. The rocks contain abundant remains of plants. This is an association of channel deposits, the most coarse and cross-bedded rocks, with facies of shoal and beach marine zones. The latter type is represented by siltstones and fine well-sorted sands with wavy bedding. Silts form thick members gently truncating one another. One of the characteristic features is the diagenetic dolomite cementation in sands.
>
> Submarine deltoid facies are represented by sediments of the delta slope, i.e., its edge most advanced toward the sea, and by those deposited on the upper delta plateau above the sea-slope deposits.
>
> The slope deposits are monotonous, vaguely laminated clayey–silty sands. They are poorly sorted, and the term *sands* is conventionally used here, because their silty and pelitic fractions are commensurable in volume with the sandy ones. The plant detritus is abundant. Fossils include the remains of marine fish (sharks), mammalian bones (Grossgeim, 1960), sponge spicules, and foraminifers. These rocks occur as a member of about 11 m in thickness below the manganese-bearing deposits.

The prospecting drilling showed more precisely that contours of paleodelta are located between the Belaya and Laba rivers (Fig. 4). Without a doubt, all these coarse clastic rocks are products deposited by a river of the Oligocene-Miocene time, which drained the western part of the Caucasus Island.

Above, we mentioned the regions of the Dagestan Mountains, where the Chokrak deposits directly rest on the Upper Cretaceous limestones. These relics of the Middle Miocene rocks occurring at high watersheds as a result of occasional preservation were described by Drobyshev (1925) and Vassoevich (1959) in the Darada, Butsra, Akushi, and other localities of the Khadzhal-Makhan anticline. Such sections are of special interest, because the Paleocene, Eocene, and par-

tially, the Maikop and Tarkhan deposits are missing here. Any submarine erosion was hardly responsible for such a significant hiatus in sedimentation. It is more reasonable to relate this hiatus with the expansion of the Caucasus territory, which was already above sea level during the Oligocene-early Miocene time and subjected to an intense denudation (Milanovsky, 1968).

The general facial plan of the Oligocene–Early Miocene deposits and spatial distribution of their terrigenous mineralogical provinces are also concordant with the existence of the Caucasus landmass during the respective period.

Approximately during the Middle Oligocene, the western and eastern parts of Ciscaucasia were regions of deposition of coarse sands and conglomerates. These rocks display features of soft-sediment deformations, and locally they grade into lumpy avalanche facies. Schatsky (1929), as well as Vassoevich and Korotkov (1935), reported on such landslide and avalanche formations from northern Dagestan, the Chernye Mountains, and other areas of the northwestern Caucasus, where transportation of the material was often oriented from the south to the north, and the coastal coarse facies graded into fine-grained clayey deposits following the same direction. Many researchers believe that this gradation indicated a rapid rising of the coastal zones of the Caucasus landmass that was associated with coastal avalanches, submarine landslides, and earthquakes. The Late Oligocene–Early Miocene time was presumably the cessation period for these tectonic movements, when sandy and silty deposits bordering the Caucasus landmass were separated from platform-derived sands by a wide zone of clayey sediments.

After his detailed study on terrigenous mineralogy and indicators of marine currents in the Maikop basin, Grossgeim (1959, p. 68) wrote:

> Northern sands and siltstones are definitely of platform origin (quartz, disthene, staurolite, sillimanite, zoisite, ilmenite), and their material was transported from inner platform areas. The southern material is, beyond a doubt, of Caucasus origin. Because of a deep erosion, the provenance here was enriched in the widespread clayey-sandy deposits of the Jurassic. The surprising similarity of the mineralogical composition of the Jurassic and Maikop deposits (abundant chloritoid, apatite, etc.) should be especially emphasized. One more point is also important. Both the siltstones and clays of the Maikop Group from lithofacies of Caucasus origin display uniform mineralogical composition (naturally, when corrected for the granulometric type of sediments). As for the clay zone, its aleuritic fractions are very poor in mineralogical composition, and do not afford a possibility to clarify their origin. Admittedly, the material could be transported in this part of the basin from both the southern and northern provenances.

Fully sharing the above considerations, we would like to stress that the Russian platform provenance for a long period influenced the sedimentation processes in the Mesozoic and Cenozoic seas, which surrounded it in the south. However, accessory minerals inherited from this provenance were redeposited many times during the evolution of the Caucasus orogen, and their associations are sometimes of unclear significance. For instance, staurolite, andalusite, and disthene occurring in the Maikop deposits of the fore-Caucasus depression are incompatible with a low degree of sandy material sorting in the host sediments (Nedumov, 1993, 1994). Nevertheless, this situation is understandable if we suggest that Jurassic and Cretaceous deposits were originally formed on the base of the mature platform material, and then, after an uprise of Caucasus Island, this material was redeposited by rivers, creeks, coastal erosion, and avalanches.

In conclusion, we would like to note that dealing with the problems of paleogeography and particular geological events, in our case, with the problems concerning Caucasus Island of the Oligocene–Miocene time, we need a complex systematic approach to solve them. In other words, we may correctly clarify such paleogeographic problems only by means of data of different geological branches. From this viewpoint, we suggest that a landmass in the Caucasus region appeared first during the pre-Oligocene time. Its configuration, relief, and landscapes were repeatedly changing with time, though the general evolutionary trend was directed to intensify rising and gradually transform the original lowland into an immense orogenic structure.

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Tectonics and Stratigraphy of the Perekatnaya–Chernaya Interstream Area, Southern Sikhote-Alin

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Abstract—The geological structure of the Perekatnaya–Chernaya (Imbisha–Taukhe) interstream area is represented by a complex combination of synsedimentary tectonic inclusions (allochthonous sheets of different origins and age ranges) embedded in a turbidite-olistostrome matrix. Three tectono-stratigraphic units of this type are distinguishable within the regional section, which are amalgamated because of their accretion. Age constraints for the matrix and tectonic inclusions are based on the distribution of radiolarians.

Key words: *geological structure, transgessions, regressions, olistostromes, Early Cretaceous, radiolarians, Sikhote-Alin.*

INTRODUCTION

The Perekatnaya–Chernaya (Imbisha–Taukhe) interstream area is located in the southern Primor'e (Fig. 1). Geological investigations, including geological mapping at a scale of 1 : 50 000, were conducted here in the 1960s. Their results provided a basis for subdividing the deposits of this region into terrigenous and siliceous-terrigenous groups with wide stratigraphic ranges. The basal unit of the sedimentary section was thought to be represented by Upper Carboniferous–Lower Permian siltstones alternating with sandstones, shales, limestone lenses, and spilite interbeds (*Geologiya* ..., 1969). This age range was estimated using foraminifers from limestones that constitute, as was then believed, sedimentary lenses. The next unit of the postulated section comprised a group of Upper Permian monotonous arkose with thin siltstone interbeds enclosing chert and limestone "lenses." The latter yielded Late Permian foraminifers. The sandstone group was thought to grade upwards into Upper Permian cherts and clayey-siliceous rocks. Mesozoic, dominantly terrigenous formations were placed above the unconformity, separating them from the Paleozoic deposits. This part of the section was subdivided into two units: the Upper Triassic formation of sandstones and siltstones bearing *Monotis* and *Halobia* fauna and the Lower Cretaceous Taukhe Formation, which was also composed of sandstones and siltstones that showed flyschoid alternation. The latter formation yielded abundant fauna and flora of the Berriasian–Valanginian age. Thus, the stratified succession of deposits in the region was thought to be of a normal transgressive-regressive type, and this viewpoint corresponded well with the former general concept of the geological structure of the Sikhote-Alin.

However, over the past decade, investigations have provided new data that cardinally change the traditional understanding of the geology of the Sikhote-Alin. First, a wide distribution of olistostromes, indicators of specific geodynamic environments of sedimentation, was reported (Mel'nikov and Golozubov, 1980; Kemkin, 1985, 1989a; Golozubov and Mel'nikov, 1986; *Mikstity* ..., 1988). Second, it was shown that, in many sections, older formations occur above (and inside) younger deposits (Mazarovich, 1985; Golozubov and Mel'nikov, 1986; Yushmanov, 1986; Mikhailov *et al.*, 1987; Kemkin, 1989b). Finally, the modern concept of the accretionary origin of the Sikhote-Alin was based on all the available data (Khanchuk *et al.*, 1989, 1991; Natal'in, 1991; Kemkin and Khanchuk, 1992, 1993; Natal'in and Borukaev, 1992). According to this concept, the normal stratigraphic succession of the deposits was disregarded, and the geological structure of the Sikhote-Alin was interpreted as a complex association of genetically diverse allochthonous formations (mostly of oceanic origin) that were tectonically embedded into a matrix of accretionary prisms composed of turbidite and olistostrome deposits.

The geology of the Perekatnaya–Chernaya (Imbisha–Taukhe) interstream area has not been considered in terms of this new concept. This work represents a first attempt to provide deeper insight into the geological structure of the region and to show the relations between respective deposits and the accretionary formations.

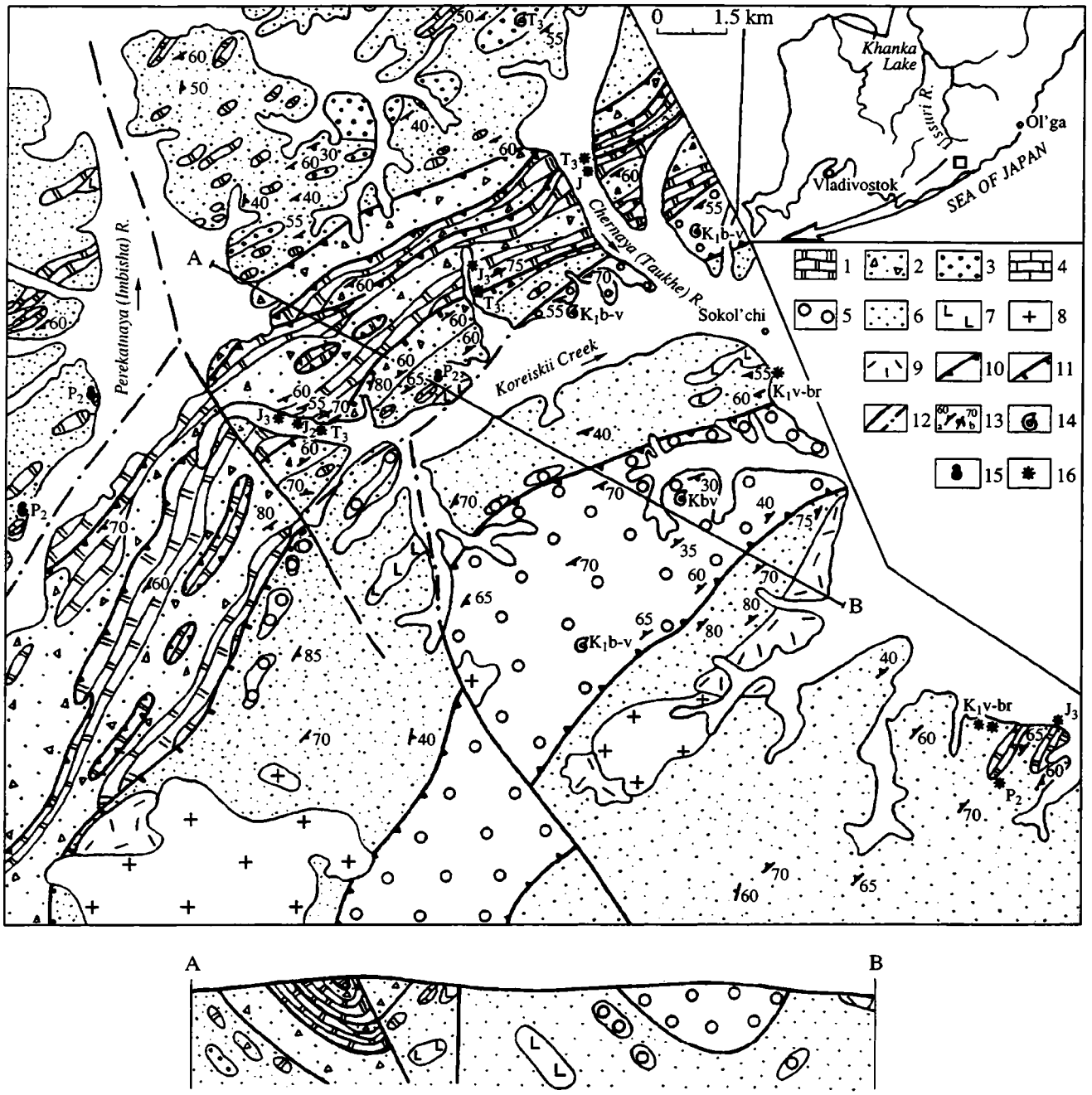

Fig. 1. Geological scheme of the Perekatnaya–Chernaya interstream area.

1—Triassic-Jurassic cherts and clayey-siliceous rocks; 2—Lower Cretaceous sandstones; 3—Triassic sandstones and siltstones; 4—Carboniferous-Permian and Permian limestones; 5—Lower Cretaceous sandstones and siltstones; 6—Lower Cretaceous turbidite-olistostrome deposits; 7—basalts; 8—Late Cretaceous granite; 9—Upper Cretaceous volcanic rocks; 10—synsedimentary thrust faults; 11—synfolding thrust faults; 12—faults (confident and implied); 13—strike and dip, normal (a) and overturned (b); 14–16—sites of sampled macrofauna (14), foraminifers (15), and radiolarians (16).

INVESTIGATION RESULTS

New data on lithology, structural geology, and paleontology indicate that the deposits in the studied region are divisible into turbidite-olistostrome, terrigenous-siliceous, and dominantly terrigenous groups.

The turbidite-olistostrome deposits form a succession of recurrently alternating turbidite and olistostrome horizons, related to each other by gradual transitions. In the stratigraphic range, this succession corresponds to the former Upper Carboniferous–Lower Permian group of siltstones and sandstones.

The olistostromes are chaotic rock associations with unsorted blocks embedded into an aleurite–psammite–argillite matrix. The blocks (olistoliths) are highly variable in size (from several millimeters up to dozens or even hundreds meters). Some of them are as large as 700–800 m across. Their form varies from isometric to irregular, sometimes elongated and lenslike. The olistoliths compose up to 35–40% of the rock volume, and they differ in their ages and compositions. Among them, I distinguished the following rock types: Carboniferous-Permian and Permian limestones; Triassic sandstones and siltstones; Triassic and Jurassic cherts and clayey-siliceous rocks; Lower Cretaceous (?) siltstones, sandstones, and gravelstones; and basalts. The ages of olistoliths are constrained by foraminifers for the carbonate rocks and by *Monotis* and *Halobia* fauna for the terrigenous deposits (*Geologiya* ..., 1969). The age of the siliceous rocks was refined on the basis of my finds of radiolarians (Plate I, specimen 6). In their lithologic features, the olistoliths of siltstones, sandstones, and gravelstones are correlative with respective terrigenous formations bearing Berriasian-Valanginian fauna and flora and with terrigenous members of terrigenous-siliceous formations. The olistostrome matrix consists of poorly sorted siltstones. It contains abundant sandy admixture and shows disharmonious microfolding indicative of slipping and flow slide of unconsolidated sediments. The thickness of the olistostrome horizons is from 10–40 to 100–200 m.

The horizons of turbidite deposits are from several to hundreds of meters in thickness. The gradual transition from the olistostrome beds to the turbidites is expressed in the progressively decreasing amount and size of the blocks embedded in rocks and in their disappearance. The turbidites consist mostly of quartz-feldspar sandstones (medium- to fine-grained) and siltstones, both of which show a poor sorting of clastic material and features characteristic of graded bedding. The bedding rhythms display progressive transitions, first from medium-grained to fine-grained sandstones, then to silty sandstones and sandy siltstones, and finally to normal siltstones and silty shales. The thickness of the cyclite elements varies from 5–7 to 25–40 cm.

The Valanginian-Hauterivian age of the turbidite-olistostrome deposits is constrained by radiolarians separated from the silty shales of the turbidite horizons (joint determinations with Rudenko; Kemkin *et al.*, 1992).

The terrigenous-siliceous rocks occur as allochthonous sheets of variable thickness within the turbidite-olistostrome deposits. In comparison to the host deposits, these rocks are older (see below). Fragments of such rocks also occur in the underlying olistostrome, indicating their synsedimentary origin, i.e., the embedding of the allochthonous sheets into the turbidite-olistostrome sequence was synchronous with the deposition of the latter. The sheets are conformable to the turbidite-olistostrome bedding, and both show a single structural style of folding. In the Koreiskii Creek headstream (Fig. 1), the deposits under consideration form a unity of four thrust sheets overlapping one another. Formerly, the terrigenous-siliceous rocks were considered as representing the Upper Permian formation of sandstones and cherts.

The terrigenous and siliceous units form an interrelated succession. The latter unit is represented by pure or clayey cherts. Their beds (from 1.5–3.0 to 7–15 cm in thickness) enclose thin (1–3 mm) laminae of siliceous shales. The rocks reveal a high degree of cataclasis; they are also deformed into disharmonious folds of low amplitudes, which are dissected by abundant fissures or microfaults. The apparent thickness of the siliceous succession is less than 100–150 m. The cherts yielded abundant radiolarians (Tables 1 and 2; Plates II and III) of the Late Triassic or the Early-Middle and Late Jurassic, and this fauna defines the respective stratigraphic ranges of the rocks. The cherts conformably grade upward into the terrigenous rocks. They are directly covered with siliceous shales overlain by siltstones with fine horizontal lamination. Above these rocks, there is a member of alternating siltstones and sandstones. The uppermost beds consist of massive, medium- to coarse-grained sandstones with acute-angled fragments of siltstones, siliceous shales, and cherts (from several mm to several cm in size). The apparent thickness of the terrigenous succession is 300–350 m. The radiolarian assemblage from the lowermost siliceous shales is similar to that separated from the underlying Upper Jurassic cherts (Table 1; Plate IV). The siltstones and sandstones are barren of fossils. Presumably, they may correspond to basal Cretaceous (Berriasian) beds, because the underlying cherts and siliceous shales yielded radiolarians of the latest Jurassic (Tithonian) age.

The dominantly terrigenous rocks also occur as synsedimentary allochthonous sheets within the turbidite-olistostrome matrix. In the basal part of the sheets, these rocks are represented by small- to medium-pebble conglomerate with gravelstone interlayers. The remaining portion of the sheet section (up to 90%) consists of finely alternating sandstones and siltstones, which bear abundant Berriasian-Valanginian fauna and flora (*Geologiya* ..., 1969). The apparent thickness of these deposits is up to 1500 m.

In addition, the turbidite-olistostrome formations enclose solitary sheets of cherts (up to 50–70 m in thickness), which yielded the Late Permian radiolarians *Follicucullus porectus* Rudenko and *Follicucullus* sp.

DISCUSSION

It is clear from the above data that, in the Perekatnaya–Chernaya (Imbisha–Taukhe) interstream area, we are dealing with three groups of geological formations that differ in age and lithology and exemplify different environments of sedimentation. For instance, the

Plate I

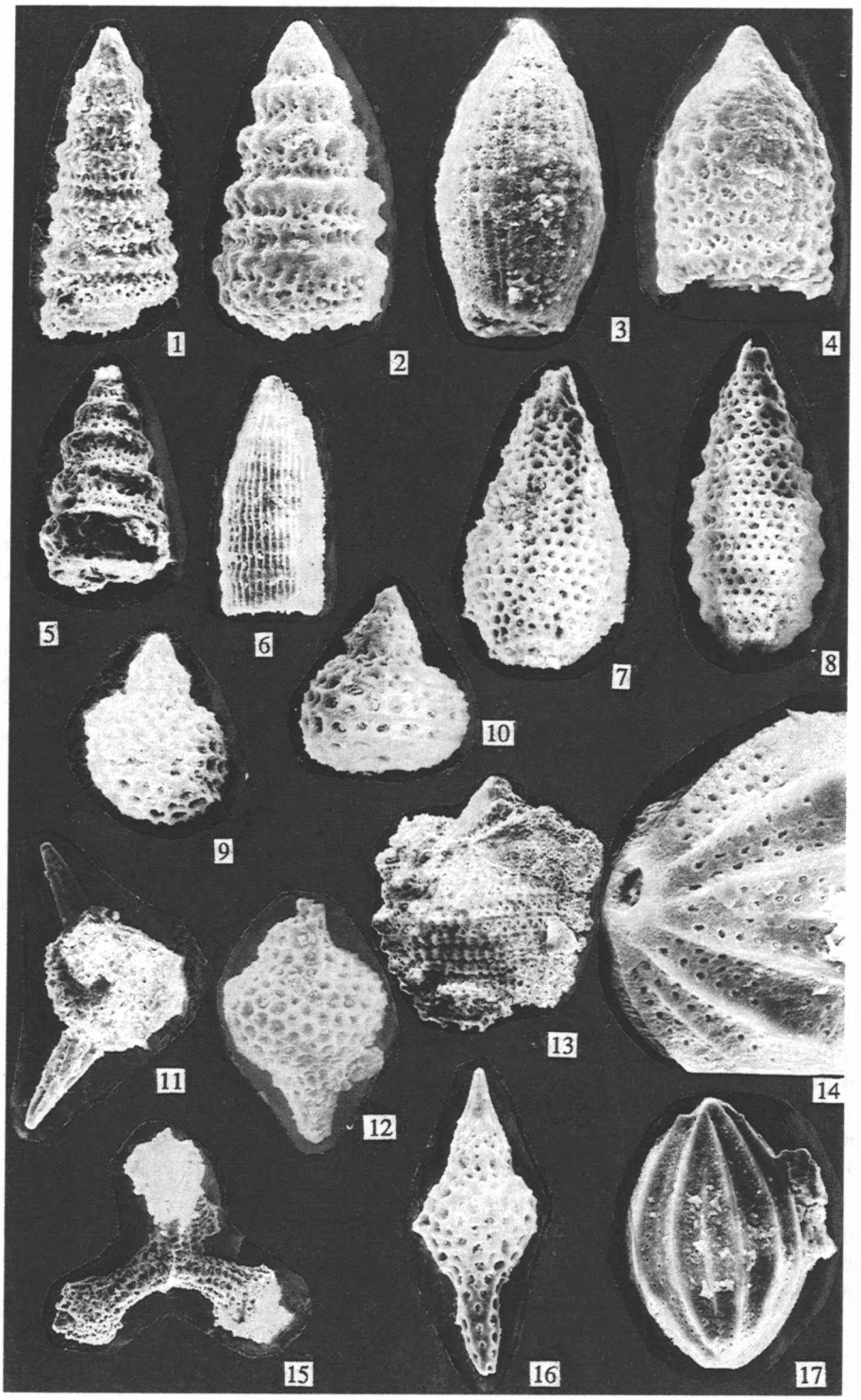

Plate I. Late Jurassic radiolarians from the cherty block embedded into the Lower Cretaceous olistostrome.

1, 2, 5—*Cinguloturris carpatica* Dumitrica [11][1]: 1—×165, 2—×250, 5—×165; 3—*Thanarla* aff. *conica* (Aliev), [11], ×250; 4—*Ristola altissima* (Rust), [11], ×165; 6—*Archaeodictyomitra* cf. *rigida* (Pessagno), [11], ×165; 7—*Amphipyndax* aff. *stoki* (Campbell et Clark), [11], ×250; 8—*Parvicingula* cf. *dhimenaensis* Baumgartner, [11], ×250; 9, 10—*Sethocapsa* cf. *uterculus* (Parona), [11]: 9—×165, 10—×250; 11—*Triactoma* cf. *hybum* Foreman, [11], ×110; 12, 16—*Podobursa triacantha* (Fischli), [11], ×165; 13—*Mirifusus* cf. *mediodilatatus* (Rust), [11], ×90; 14, 17—*Protunuma ochiensis* Matsuoka, [11]: 14—×500, 17—×250; 15—*Paronaella mulleri* Pessagno, [11], ×90.

[1] From here on, figures in brackets indicate the specimen number in the Tables.

Table 1. Late Jurassic radiolarians from cherts and shales of the terrigenous-siliceous allochthons and olistolith

Species	Samples										
	1	2	3	4	5	6	7	8	9	10	11
Thanarla conica (Aliev)	*	*	*	*		*	*	*			*
T. pulchra (Squinabol)									*	*	
Archaeodictyomitra rigida Pessagno			*	*			*	*			*
A. apiara (Rust)			*	*					*	*	
A. ex gr. *apiara* Pessagno	*			*							
Archaeodictyomitra sp.									*	*	
Tritrabs cf. *ewingi* Pessagno	*							*		*	
Paronaella mulleri Pessagno									*	*	*
Paronaella sp.									*	*	
Tricolocapsa rusti Tan Sin Hok	*			*	*	*		*			
Podobursa triacantha (Fischli)									*	*	*
P. polylophia Foreman										*	
Podobursa sp.			*								
Sethocapsa uterculus (Parona)					*		*	*	*	*	*
S. cetia Foreman								*			
Sethocapsa sp.			*	*			*				
Zhamoidellum ovum Dumitrica	*		*								
Xitus cf. *plenus* Pessagno	*		*		*		*	*			
X. spicularis (Aliev)	*		*	*			*		*	*	
Pseudodictyomitra minoensis Mizutani	*		*	*							
P. leptoconica (Foreman)	*	*					*				
P. carpatica (Lozynyak)	*	*	*	*	*	*		*	*	*	
Mita sp.			*								
Ristola altissima (Rust)	*		*	*					*	*	*
Amphipindax plousies Foreman									*		
Amhpypindax sp.											*
Protununa fusiformis Ishikawa & Yao			*	*							
P. ochiensis Matsuoka											*
Crucella theokaftensis Baumgartner										*	
Triactoma cornuta Baumgartner									*	*	
T. aff. *hybum* Foreman											*
Mirifusus mediodilatatus (Rust)									*	*	*
Holocryptocanium geysersense Pessagno			*					*			
H. barbi Dumitrica									*		
Cinguloturis carpatica Dumitrica				*					*	*	*
C. cylingra Kemkin et Rudenko	*										
Eucyrtidium cf. *unumaensis* Yao					*						
Parvicingula mashitaensis Mizutani	*									*	
P. cf. *dhimenaensis* Baumgartner											*
P. boesi (Parona)	*			*	*						
Archaeospongoprunum imlayi Pessagno	*								*	*	
Stichocapsa altiforamina Tumanda								*			
Stichocapsa sp.									*		
Pseudoaulophacus sp.										*	

Note: That column numbers correspond to the following sample indexes and rock types: 1*—T-22**, 2—T-23, 3—T-28, 4—T-33, 5—T-34, 6—T-35, 7—T-36, 8—T-38, cherts; 9—T-45, 10—T-46, shales; 11—T-47, olistolith (*—number in Tables 1 and 2; **—sample number).

Table 2. Late Triassic and Early-Middle Jurassic radiolarians from cherts of the terrigenous-siliceous allochthons and olistolith

Species	Samples									
	12	13	14	15	16	17	18	19	20	21
Triassocampe sp.	*	*	*	*	*	*	*			
Canoptum sp.	*	*	*				*			
Capnodoce anapetes Dewever		*								
Capnodoce sp.	*	*		*						
Tripocyclia sp.				*	*	*				
T. japonica Nakaseko et Nishimura						*				
Archaeodictyomitra rigida Pessagno								*		*
Archaeodictyomitra sp.								*	*	
Hsuum maxwelli Pessagno								*		
Hsuum sp.								*		*
Parvicingula sp.									*	*
P. elegans Pessagno et Whalen										*
Archaeospongoprunum sp.								*		
A. tehamaensis Pessagno										*
Noritus lillihornensis Pessagno et Whalen										*
Eucyrtidium sp.									*	
Paronaella sp.								*	*	
Tricolocapsa rusti Tan Sin Hok								*		

Note: That column numbers correspond to the following sample indexes and rock types: 12—T-29, 13—T-31, 14—T-32, 15—T-40, 16—T-42, 17—T-3, 19—T-10, 20—T-17, 21—T-26, cherts; 18—T-48, olistolith.

coarse terrigenous rocks with abundant flora and macrofauna were evidently deposited in shallow settings, while cherts of the terrigenous-siliceous group are typical of pelagic environments within areas remote from any provenance of terrigenous material. The olistostromes represent specific sedimentary facies of submarine slopes of the convergent margins associated with subduction zones (Ueda, 1983; Hilde, 1983; *A moins* ..., 1988). Evidently, the allochthonous synsedimentary sheets of rocks that differ in age and lithology could have been embedded into the turbidite-olistostrome matrix only after their considerable displacement from the original settings, followed by accretion. Disregarding details and the possible scales of these processes, I will only note that they are recorded in the following features: first, in the gradual replacement of siliceous sedimentation by the deposition of clastic rocks, which were initially fine-grained, and then, as the accretion zone was approached, consisted of coarse terrigenous material; and second, in the occurrence of such chaotic formations as olistostromes deposited during the accretion of allochthonous sheets.

Thus, the studied complex of deposits definitely consists of turbidite-olistostrome matrix and embedded synsedimentary allochthons. In other words, these formations constitute a fragment of the Early Cretaceous accretionary wedge, consisting of three tectono-stratigraphic units amalgamated into a single section as a result of accretion (Fig. 2).

Data on the composition, structure, and age of the allochthonous sheets and matrix show that this ensemble of rocks is correlative with the Early Cretaceous accretionary complex in northern regions (Dal'negorsk and Kavalerovo) of the Sikhote-Alin (Khanchuk *et al.*, 1989, 1991). In these regions, the Lower Cretaceous turbidite-olistostrome deposits contain olistoliths of the following rocks: Triassic–Jurassic cherts with overlying terrigenous sediments of the Lower Cretaceous (Berriasian), Permian and Triassic limestones sometimes associated with alkaline basalts (fragments of

Plate II

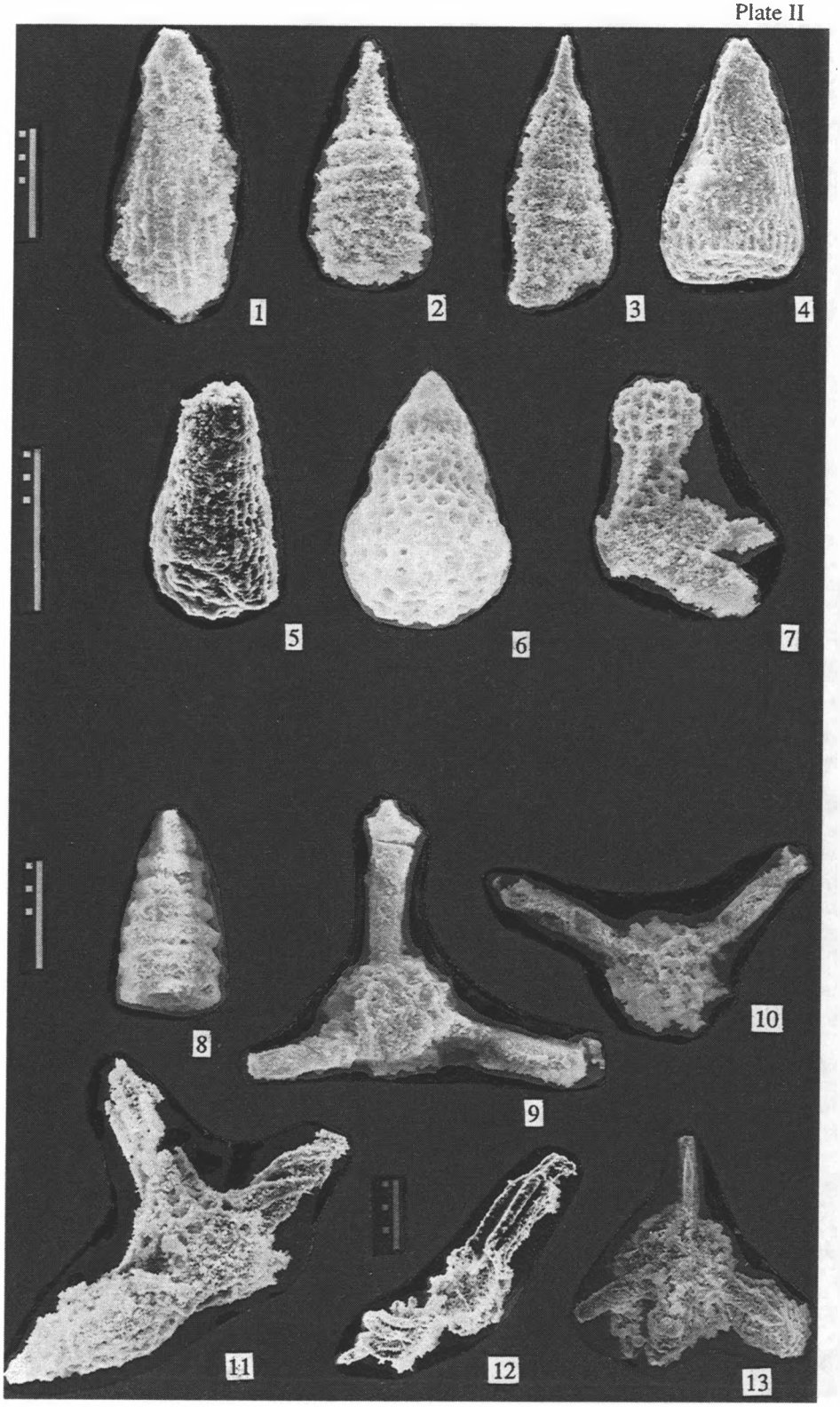

Plate II. Late Triassic and Early–Middle Jurassic radiolarians from siliceous allochthons.

1—*Archaeodictyomitra* cf. *rigida* (Pessagno), [21]; 2—*Parvicingula* aff. *elegans* Pessagno et Whalen, [21]; 3—*Noritus lillihornensis* Pessagno et Whalen, [21]; 4—*Hsuum maxwelli* Pessagno, [19]; 5—*Hsuum* sp., [19]; 6—*Eucyrtidium* sp., [20]; 7—*Paronaella* sp., [20]; 8—*Triassocampe* sp., [13]; 9, 10—*Capnodoce anapetes* Dewever, [13]; 11—*Tripocyclia* cf. *japonica* Nakaseko et Nishimura, [17]; 12—*Tripocyclia* sp., [17]; 13—*Capnodoce* sp., [12]; bar = 100 μm.

Plate III

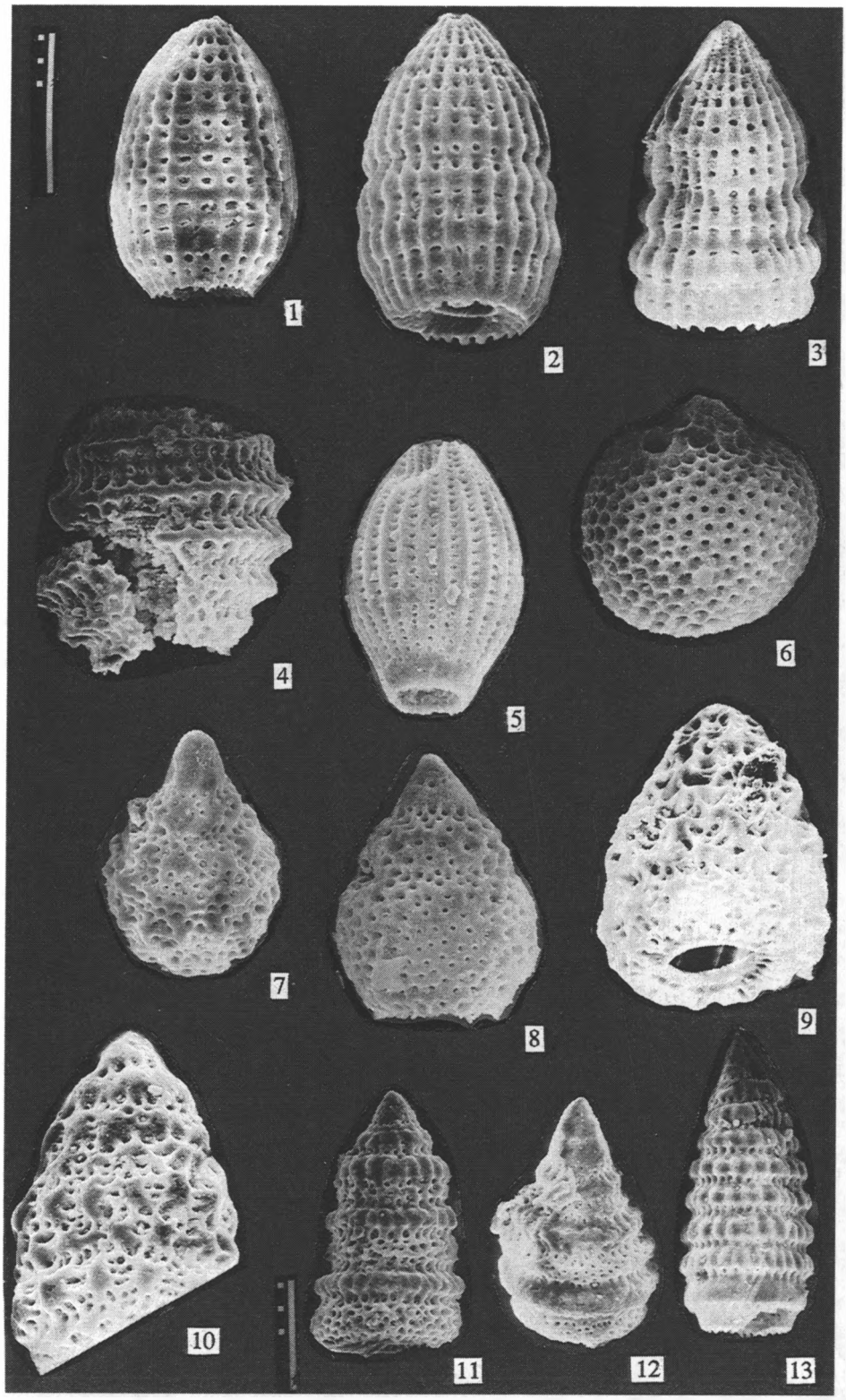

Plate III. Late Jurassic radiolarians from siliceous allochthons.

1—*Archaeodictyomitra apiara* (Rust), [4]; 2, 3—*Archaeodictyomitra* ex gr. *apiara* (Rust), [1, 4]; 4—*Parvicingula* cf. *mashitaensis* Mizutani, [1]; 5—*Thanarla* cf. *conica* (Aliev), [4]; 6—*Zhamoidellum ovum* Dumitrica, [1]; 7—*Sethocapsa cetia* Foreman, [8]; 8—*Stichocapsa* cf. *altiforamina* Tumanda, [8]; 9, 10—*Xitus plenus* Pessagno, [7, 5]; 11—*Cinguloturris cylindra* Kemkin et Rudenko, [1]; 12—*Cinguloturris* sp., [4]; 13—*Pseudodictyomitra leptoconica* (Foreman), [1]; bar = 100 μm.

Plate IV

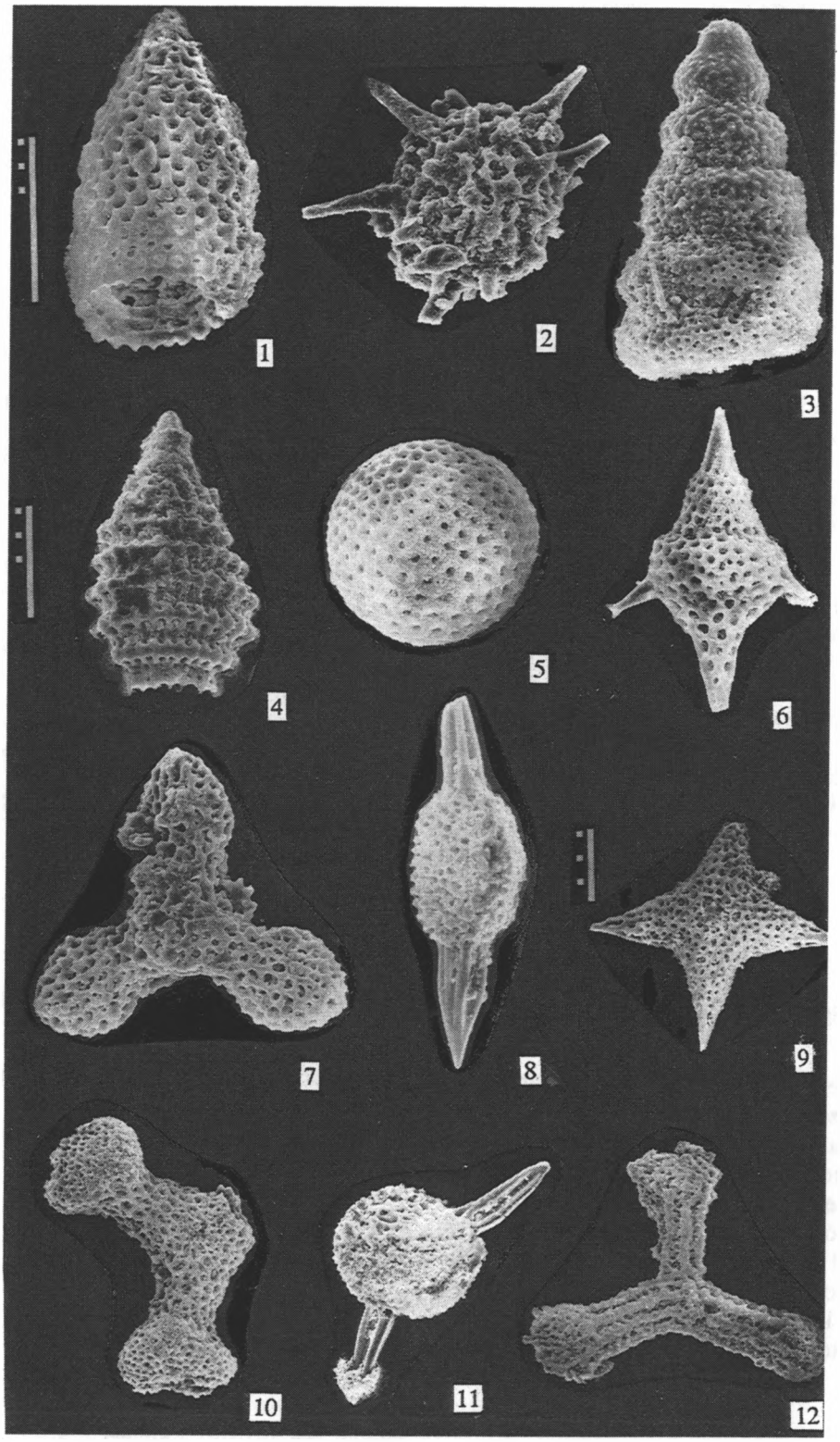

Plate IV. Late Jurassic radiolarians from shales of terrigenous-siliceous allochthons.
1—*Stichocapsa* sp., [9]; 2—*Pseudoaulophacus* sp., [10]; 3—*Amphipyndax plousies* Foreman, [9]; 4—*Parvicingula mashitaensis* Mizutani, [10]; 5—*Holocryptocanium barbi* Dumitrica, [9]; 6—*Podobursa triacantha* (Fischli), [10]; 7—*Paronaella mulleri* Pessagno, [9]; 8—*Archaeospongoprunum imlayi* Pessagno, [10]; 9—*Crucella theokaftensis* Baumgartner, [10]; 10—*Paronaella* sp., [9]; 11—*Triactoma cornuta* Baumgartner, [9]; 12—*Tritrabs* cf. *ewingi* (Pessagno), [10]; bar = 100 μm.

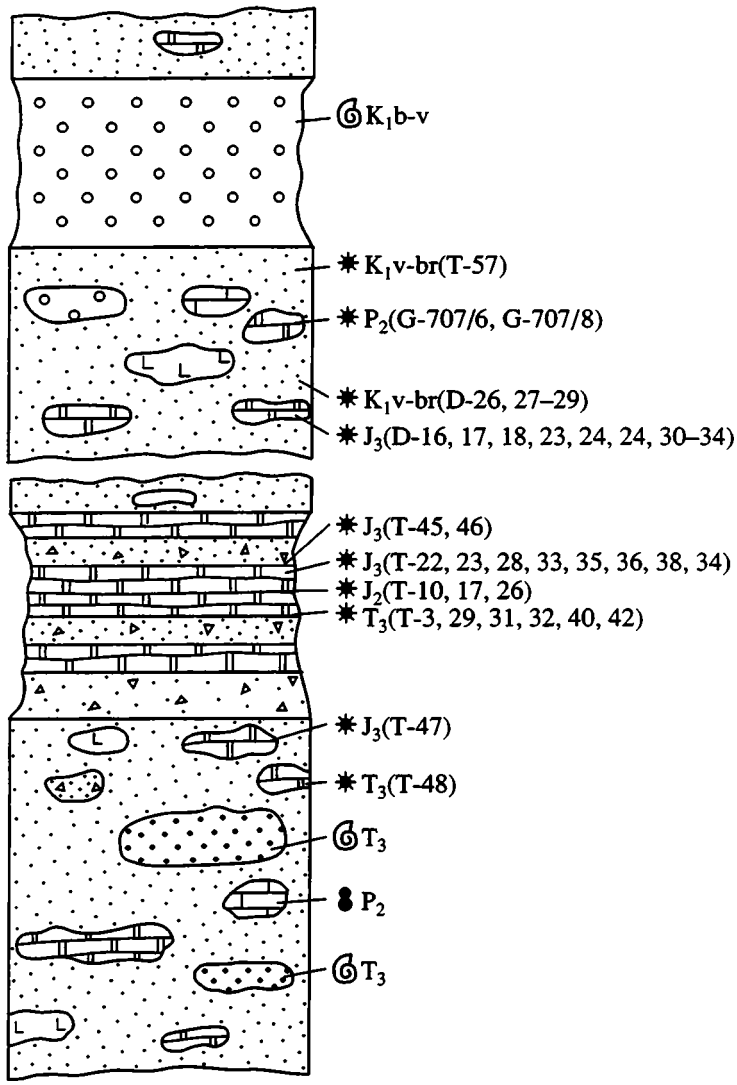

Fig. 2. Tectono-stratigraphic column of deposits in the Perekatnaya–Chernaya (Imbisha–Taukhe) interstream area (symbols as in Fig. 1).

past guyots), coarse clastic sediments of the Early Cretaceous age, Triassic terrigenous rocks with *Monotis* and *Halobia* fauna (*Mikstity* ..., 1988), and Permian cherts. In the region of the current study, the large sheets of limestones are absent, although their small blocks have been observed in the olistostromes. This seems to be related to structural variations of the Early Cretaceous accretionary complex in different regions. Blocks of Triassic limestones are also unknown, probably because the carbonate rocks of the region are inadequately studied.

CONCLUSION

1. In the Perekatnaya–Chernaya (Imbisha–Taukhe) interstream area, a fragment of the Early Cretaceous accretionary wedge is exposed, which consists of three tectono-stratigraphic units incorporated into a joint structure.

2. The obtained data on the composition, structure, and age of the sedimentary matrix and enclosed allochthonous sheets of the wedge are correlative with respective data for adjacent regions, and they support the concept that the Early Cretaceous accretionary complex is traceable all along the coastal zone of the Sikhote-Alin.

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Paleogeographic Interpretation of Paleoflora from the Pliocene–Pleistocene Sediments of Western Syria

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Abstract—Complex analysis of paleoflora was used to study the Mardido reference section (northwestern coast of Syria). It was found that this section was formed in coastal and sometimes estuarine settings. Palynological data allowed four alternating humid and arid periods to be distinguished. Six intervals of salinity oscillations of the basin water regime were recognized by the analysis of diatom algae and nannoflora. Short-range marine invasions were depicted against a background of a general tendency of freshening and shallowing of the basin. In general, the variation trends of species composition for the aquatic and continental floras were similar, thus providing a basis for the correlation of geological events.

Keywords: *paleogeography, Pliocene, Pleistocene, complex analysis, nannoplankton, diatoms, palynology.*

E.V. Devyatkin, A.E. Dodonov, K. Khatib, and Kh. Nseir were part of a group of scientists who was the first to study the marine Pliocene–Pleistocene deposits of western Syria in detail. The work was carried out by a joint Soviet-Syrian expedition during 1990–1992. A detailed description of the paleomagnetic and biostratigraphic results as well as the geological structure of the Mardido, Msherpheh, Al Quatriya, and Jinndiriye sections were given in reports of the expedition and in an article published in this volume (Devyatkin *et al.*, 1996). A study of the influence of the Pliocene–Quaternary climatic fluctuations on assemblages of continental and marine flora and a reconstruction of the paleogeographic environments in the Latakia depression (using the Mardido section as an example) are goals of the present work. To solve paleogeographic problems, the relative reaction rates of various floral groups, both continental and aquatic, to global climatic changes must be known. Data on the correlation between events occurring in water and on land are desirable in order to choose a correct approach to solving problems of stratigraphy and the correlation of geological events.

Due to the rather specific environmental conditions of flora existence, the coastal-marine sediments of the Mardido section include impoverished assemblages of marine planktonic organisms. Periodic salinity variations caused by an increase (or decrease) in river runoff as well as progressive shallowing of the basin resulted in a discrete distribution of nannoflora and diatoms. From the bottom upward in the section, nannoplanktonic assemblages become progressively more impoverished; only the most ecologically flexible species survive up to the topmost part.

Five intervals of the section corresponding to certain paleogeographic conditions are described below (figure).

I. 0–5 m

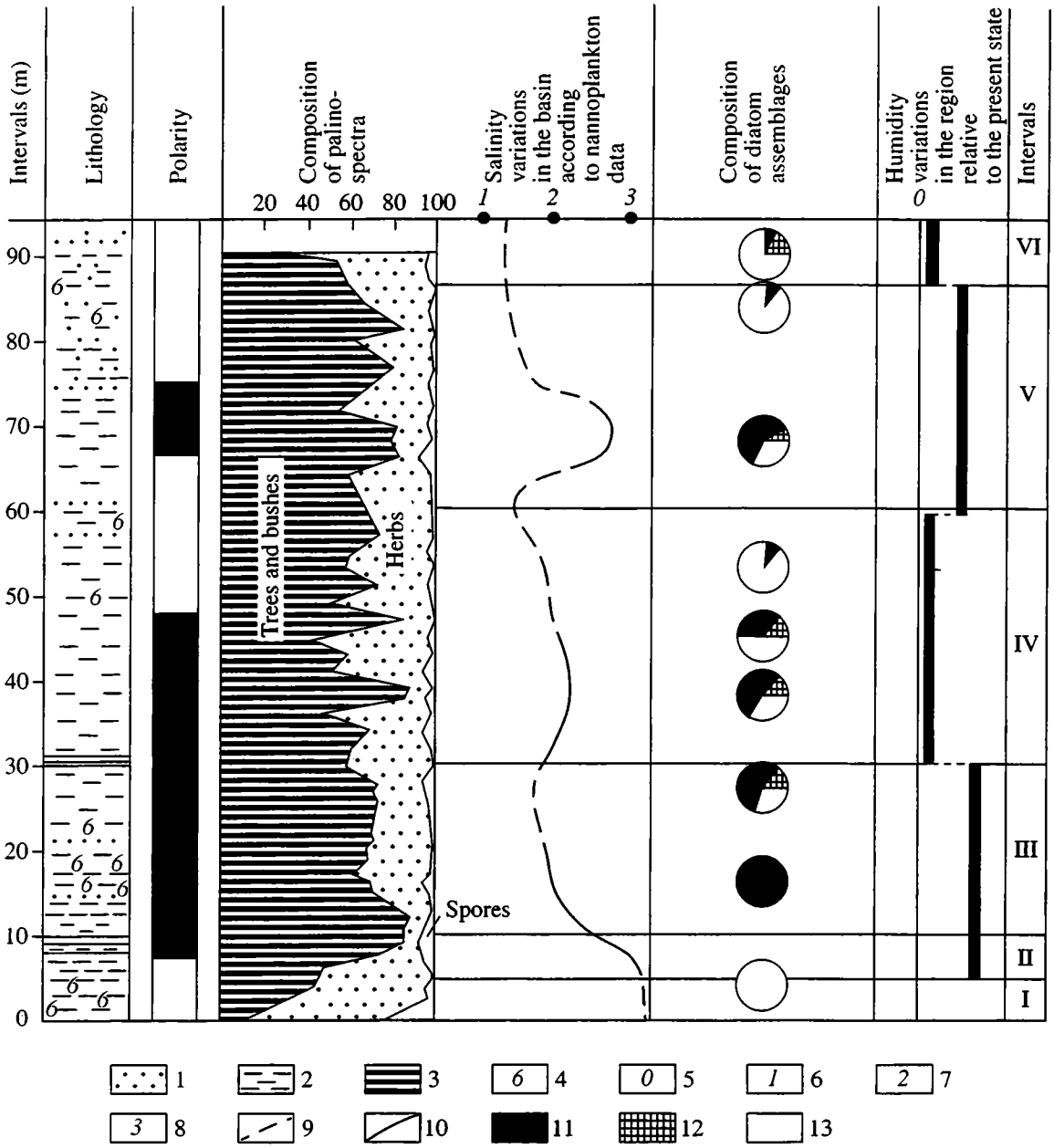
Herbaceous pollen, mainly that of composite plants, predominates in the spore and pollen spectra. An admixture of modern pollen is suspected, and this results in a significant distortion of the autochthonous spectra.

Nannoplankton was not found in this interval. A rather poor diatom assemblage of exclusively freshwater species was found at a depth of 4 m. Soil-dwelling forms *Pinnularia borealis* et var. *brevicostata* and *Hantzschia amphioxys* predominate in it. All of this, together with the presence of the bog species *Pinnularia viridis*, probably indicates the existence of a strongly freshened shallow basin at that time.

II. 6–10 m

Spore and pollen spectra from this interval are rich in arboreal pollen. Coniferous pollen predominates. Corresponding paleolandscapes resembled a modern wooded savanna. Mountains were presumably covered by pine forests with spruce and fir (Simakova, 1994). Climatic conditions corresponded to a warm climate with a significant amount of seasonal precipitation.

A rich and diverse nannoflora assemblage of a relatively deep basin with normal salinity is typical of this interval. The presence of discoasters and scyphosphers, which are typical representatives of tropical and sub-



Complex analysis of paleoflora from the Mardido Section (western Syria).

1—sand; 2—loam; 3—clay; 4—molluscan shells; 5—modern state of climate; data on nannoplankton: 6—freshened conditions; 7—marine brackish-water conditions; 8—normal marine conditions; 9—absence of nannoplankton; 10—presence of nannoplankton; types of diatom assemblages: 11—marine; 12—brackish-water; 13—freshwater.

tropical nannofloras, indicates high water temperatures. Such stratigraphically important species as *Discoaster peutaratialus, D. brouweri*, and *D. surculus* are important age-markers of the deposits, and are useful in interpreting paleomagnetic data. According to paleomagnetic analysis, the Gauss–Matuyama paleomagnetic boundary is traced at the level of 8 m.

III. 10–30 m

During the period corresponding to this interval, the vegetation of coniferous broad-leaved forests, which indicates a mild and relatively cool climate, predominated on land areas.

Nannoflora from the lower part of the interval (10–18 m) is characteristic of the coastal shallow shelf area. Discoasters and scyphosphers disappeared completely. This was probably caused by both a shallowing of the basin and a decrease in temperatures; only the most resistant species, such as *Pseudoemiliania lacunosa, Helicosphaera corteri*, and *Syracosphaera* sp., survived these changes.

A large amount of precipitation probably increased the river runoff, thus steadily influencing the salinity regime of the coastal shallow sea of that time. Under progressive shallowing and freshening of the basin, nannoflora nearly disappears in the upper part of the interval (20–30 m). Diatoms from the lower part of the interval (16 m) are represented by single coastal-marine and marine brackish-water forms associated with some purely brackish-water and freshwater specimens.

IV. 30–60 m

This interval shows frequent and well-manifested variations in the general composition of spectra against a background of the decreasing role of arboreal pollen. The inferable landscapes were similar to present savannas. Pine-oak forests with spruce were probably typical of high mountains, while low mountains were covered by sclerophyllous evergreen forests. A possible significant aridization of the climate probably resulted in a decrease in river runoff. This caused a short-term increase in the influence of the sea on biota against a background of general freshening of the basin. Tectonic factors could also be responsible for this.

Nannoflora again appears in sediments of the Mardido section in the interval of 30–45 m. Its impoverished assemblage, stunted in character, suggests shallow coastal settings of sedimentation that were rich in terrigenous material. *P. lacunosa, C. macintyrei*, and *H. carteri* play the main role in this assemblage.

Nannoplankton is absent in the upper part of the interval.

The evolution of the diatom assemblages within this interval suggests further freshening of the basin, which was characteristic of the previous stage. Diatoms show extensive taxonomic diversity. Neritic and oceanic species *Thalassiosira angustelineata, T. oestrupii,* and others predominate in the lower part of the interval (30 m). The share of brackish-water elements such as *Thalassiosira decipiens* and *Cyclotella caspia*, which are typical of the present Black and Caspian seas (Proshkina-Lavrenko, 1955), increases upwards (45 m). In general, planktonic species, both marine and freshwater, predominate in the interval. The ubiquists *Stephanodiscus rotula* and *S. minutulus*, which are also common in the present Black and Mediterranean seas, are most remarkable among the freshwater forms. In the upper part of the interval (53 m), further shallowing and freshening of the paleobasin caused an increase in the role of freshwater benthic species, which are mainly represented by periphytic and lacustrine-bog forms.

V. 60–86 m

The amount of arboreal pollen increases in this interval; it is represented by single grains of lime, magnolia, and beech associated with dominant pollen of coniferous plants. This was probably a period in which sclerophyllous forests and bushes prevailed in the plains. Mountain vegetation was represented by pine forests with an admixture of spruce and fir. There was a mountain belt of deciduous species: lime, alder, birch, and beech. The climate became wetter and relatively cooler. Above the level of 66 m, the influence of marine waters increases again. A rather abundant and diverse nannoflora assemblage is found here. Triradiate forms of *D. brouweri* (*D. triradiatus*) and single specimens of *Gephyrocapsa* s. l. were found at this level, thus allowing, together with paleomagnetic data, the Pliocene–Pleistocene boundary to be traced (Rio, 1991).

The short period of increased marine influence at the Pliocene–Pleistocene boundary covers a very small interval (66–72 m). It is overlain by a sandy bed (probably, marking latent erosion) and a silty clay member, both of which are nearly barren of nannoplankton.

Similar to the nannoplankton the diatoms indicate environmental changes at this time. An algae assemblage, similar to that of interval III, was found again. Marine neritic and littoral species predominate in it; the semibenthic species *Paralia sulcata* is the most abundant. Mainly halophilic forms such as *Nitzschia capitellata, N. frustulum*, and others were distinguished among freshwater species.

VI. 86–89 m

The interval of 86–89 m revealed an increased content of herbaceous pollen (Compositae and Cruciferae) and the disappearance of fir and cedar pollen. Pollen of *Juglans, Erica, Ostrya*, and *Ilex* is present here. Typical Mediterranean xerophyte vegetation seems to have existed during the respective period. The plains were occupied by maquises, and the mountains were covered by pine-oak forests. The climate became arid, similar to the present one.

Nannoplankton was not found in this interval, but diatoms showed another change in their composition. The number of marine species sharply decreased. They became mainly represented by benthic forms, and the share of brackish-water species (*Nitzschia sigma* and *N. spectabilis*) slightly increased. Freshwater benthos plays the main role in this assemblage. It includes representatives of different ecological groups composed of bog and lacustrine-bog forms (*Eunotia linaris, Pinnularia subcapitata*, and *P. viridis* et var. *leptogongyla*), soil-dwellers (*Navicula mutica* var. *cohnii, Pinnularia borealis* et var. *brevicostata*, and *Hantzschia amphioxys*), and fluvial species (*Navicula cryptocephala* and *N. minima*). This composition of the assemblage indicates a shallow, considerably freshened basin, which still had connections with the sea. It was probably a swamped and periodically drained estuary.

CONCLUSION

Assemblages of nannoplankton, diatom algae, pollen, and spores of the reference Mardido section, situated on the northwestern coast of Syria (Latakia depression), were subjected to analysis, which led to the following paleogeographic conclusions.

According to the revealed specific features, this section was formed under coastal and sometimes estuarine conditions, in which the sea level and salinity variations in the basin were mainly affected by local factors such as fluctuations in river runoff against a background of global climatic changes and neotectonic movements.

Four alternating humid and arid periods and six intervals of salinity variations in the basin were distinguished by the analysis of diatom and nannoflora assemblages.

Periods of maximum nannoflora development (typical marine indicators) corresponded to transgressive cycles. Diatoms enabled marine and brackish-water or nearly freshwater conditions to be recognized, which correspond to regressive cycles.

Two large transgressive maxima (stages II and V) recorded in the Mardido section are well-correlative with the data obtained by the previous study on sea-level oscillations in the global ocean and the Mediterranean Sea (Fairbridge, 1972; Stanlery, 1971; Vail *et al.*, 1977; Poag, 1984; Shakhud, 1993). Respectively, they correspond to the end of the early Pliocene and to the latest Pliocene–earliest Pleistocene.

The late Pliocene regressive stage is manifested in intervals III and IV. The period of marine brackish-water conditions (at the beginning of interval IV) suggests a general tendency of climatic aridization against a background of general regression. This was caused by local factors (probably a decrease in continental river runoff), but not by global climatic changes.

Interval V, remarkable because of the slightly increased influence of the sea at the Pliocene–Pleistocene boundary, corresponds to the Calabrian transgressive phase (Emilian) of the Mediterranean or the Gurian phase of the Black and Caspian sea areas (Shakhud, 1993). A rise in the sea level of the Atlantic is also evident at the Pliocene–Pleistocene boundary (Poag, 1994; Vail *et al.*, 1977).

Thus, against the background of a general tendency of freshening and shallowing of the basin, some short-range marine invasions were distinguished. They are clearly recorded by assemblages of nannoflora and diatoms.

In general, the evolutionary trends of aquatic and continental floras were similar, which is important for the correlation of geological events. For instance, in general, marine transgressions on the western Syrian coast coincide with epochs of increases in forest areas and hydrophilous vegetation on land, which indicate humidization of climate. General regressions correspond to the maxima of steppe vegetation on land, reflecting the periods of climate aridization.

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BRIEF COMMUNICATIONS

Paleogeographic Environments in the Northwestern Regions of the Black Sea Coast in the Early Pliocene

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Key words: *Moldova, Ukraine, early Pliocene, mammals, reptiles, paleogeography.*

After the regression of the early Pontian Sea, the whole area of Moldova and southwestern Ukraine faced the onset of continental sedimentation. In this region, deposits of Pliocene fluviatile plains were recognized in three areas in the interfluve of the Prut and South Bug rivers: the Kuchurgan, Stol'nicheny, and Karboliya areas (Bilinkis, 1987). Until recently, researchers have had no uniform view on the age interrelationships of these deposits and their position in the stratigraphic scale of the Eastern Paratethys. We studied a number of sections in the valley of the Kuchurgan River, the interfluve of the Dniester and Prut rivers, the valley of Bol'shaya Salcha River, and in the vicinity of Odessa. We also partly revised materials on the mammal fauna that were collected in these areas and are stored in the Geological Institute of the Russian Academy of Sciences and used data from the literature. Our study showed a different age for these deposits. The Stol'nicheny fluviatile deposits were formed in the first half of the early Cimmerian, the Kuchurgan deposits are dated at the terminal early and middle Cimmerian, and the Karboliya deposits are dated at the late Cimmerian.

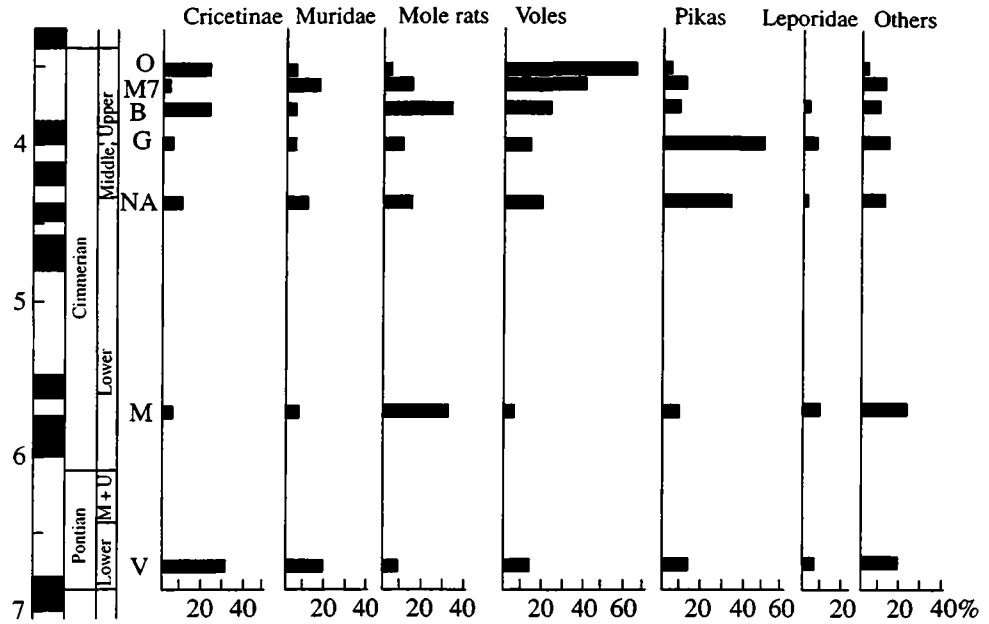
Geological information and vertebrate fauna from the lower Pontian deposits and the above-listed fluviatile members make it possible to draw certain conclusions about the development of the continental ecosystem in the northwestern Black Sea region from 7 to 3.5 Ma ago.

The composition of the early Pontian assemblage of small mammals suggests that the considered area was characterized by mosaic landscapes at that time and populated by their different ecological groups. The assemblage recorded in the localities of the Odessa–16 Station of the Bol'shoi Fontan (Topachevsky *et al.*, 1988) and the Vinogradovka locality of the Bolgrad area of the Odessa region, Ukraine (our collections), is dominated by hamsters (figure), dwellers of meadow and steppe biotopes. The presence of Gliridae, flying squirrels, and diverse Muridae (three species) indicate substantial development of arboreal vegetation, at least along the rivers. Riparial biotopes were inhabited by freshwater turtles of the *Chelydropsis* and *Sakya* genera (Chkhikvadze, 1988), beavers, and desmans. The fauna of large mammals is inferred only from scarce finds in lower Pontian limestones (Dubrovo and Kapelist, 1979; Korotkevich, 1988). It included giraffes and camels, dwellers of savanna and dry open landscapes. Mastodonts, deinotheriums, rhinos, and microstonyxes inhabited brushwoods along rivers. The occurrence of camels and the complete absence of *Tragocerines*, which were typical for the Meotian, may indicate that climatic conditions were more arid during the Pontian than the Meotian. The presence of subtropical plants in the flora (Shchekina, 1966) points to a relatively high average annual temperature.

There are almost no paleontological data on the environments of the second half of the Pontian and the beginning of the Cimmerian. The early Cimmerian situation can be inferred from scarce finds in the Stol'nicheny fluviatile deposits (Mugureny, Sagaidak, and Tritskoe localities). The fauna of small mammals of the Mugureny locality is characterized by taxonomic diversity; the different ecological groups are represented in more or less equal proportions, with the exception of some predominance of Spalacidae. The fauna of large mammals includes a small hipparion, a gazelle, a roe, and mastodonts. All these forms are typical for both open and wooded areas.

The fauna of brackish-water mollusks presents evidence that the climate in the early Cimmerian was warmer than that in the Pontian. It was also much more humid, which caused the formation of iron ore and laterite weathering crusts (Eberzin, 1940). Specifically, in the area of Odessa and along the Prut River, one can observe remnants of such a crust that developed upon the lower Pontian deposits. The time during which its formation began is presently unknown; however, most likely, it commenced in the earliest Cimmerian.

The maximum of climatic warming and, most likely, humidization occurred during the middle Cimmerian, the time of the formation of the Kuchurgan fluviatile deposits. Apparently, the activization of tectonic movement in the Carpathians caused an increase in erosional processes. This resulted in a significant coarsening of clastic materials of Carpathian origin in the



Percentage of remains of small mammals in the studied localities.
V—Vinogradovka; M—Mugureny; NA—Novaya Andriashevka; G—Grebeniki 2; B—Budei; M7—Musait 7; O—Odessa catacombs.

Kuchurgan deposits as compared to the Stol'nicheny fluviatile deposits.

The vertebrate fauna (mammals and reptiles) indicates that mosaic landscapes were preserved. The fauna of rodents and lagomorphes experienced a quantitative restructuring; the dominating role was gained by picas (Novaya Andriashevka and Grebeniki 2 localities). This was also noted by Topachevsky and Nesin (1989). Large mammals show that the dominant role was played by muntjacs of the genera *Muntiacus*, *Eostyloceros*, and *Paracervulus* together with small deers of the *Croizetoceros* and *Pliocervus* genera. All these forms were confined to biotopes with brush and arboreal vegetation situated along rivers. The same biotopes were inhabited by tapirs (unknown in older faunas), pigs *Propotamochoerus*, and mastodonts *Zygolophodon borsoni*. Primates *Macaca* and *Dolychopithecus* (Dubrovo and Kapelist, 1979) are also known. The extremely rare occurrences of hipparion remains and complete absence of giraffes and camels are worth noting.

In the late Cimmerian, Karboliya fluviatile deposits accumulated in the Prut area. These deposits rest on the eroded surface of a red weathering crust covering the lower Pontian deposits. This implies that the formation of the crust had terminated by the end of the middle Cimmerian. The sediment load of the fluviatile plain shifted westward, so that the Karboliya alluvium began to discharge into the Dacian Basin. The size of clastic particles increases up the section (Khubka, 1982). This indicates that, by the end of the considered time, there was an increase in the intensity of the tectonic movements and, accordingly, erosion in the Carpathians.

The main difference of the small mammalian fauna from that of the preceding stages is an abrupt decrease in the quantity of pikas. Their niches were apparently occupied by voles (Budei and Musait 7 localities). By the end of the Cimmerian, voles of the family Arvicolidae reached an absolute predominance and displayed adaptive radiation. The genus *Promimomys*, which existed since the beginning of the Pontian, was joined by genera *Pliomys* and *Dolomys* (localities of the Odessa catacombs and the lower bed of the Kotlovina).

The fauna of large mammals shows that the role of muntjacs somewhat decreased, with the only remaining genus *Paracervulus*, and that the number of roes increased (two species of *Procapreolus*). Brushwoods along rivers were still inhabited by *Propotamochoerus* and mastodonts. However, *Zygolophodon borsoni* occurred only rarely, whereas the role of *Anancus arvernensis* increased. The hippopotamus was also present (Godina and David, 1973; Korotkevich, 1988). Along with animals associated with humid wooded biotopes, the watershed plateaus were apparently inhabited by gazelles; very rare hipparions; and newly appearing camels, which reached high numbers by the end of the Cimmerian. These changes in the mammalian assemblages indicate a certain climatic aridization at that time.

During the formation of the Kuchurgan and Karboliya deposits (middle–late Cimmerian), the considered area was populated by an extremely rich reptile fauna. Turtle faunas of the Kuchurgan and Karboliya fluviatile deposits are similar. They include the snapping turtle *Chelydropsis nopcsai*, freshwater species *Melanochelis pidopličkoi* and *Sakya riabinini*, the marsh turtle *Emydoidea antiqua*, the land tortoise *Tes-*

tudo černovi, and other forms. The Karboliya deposits yielded diverse remains of snakes (this group is not studied in the Kuchergan deposits), which can be assigned to more than ten species of six genera (*Eryx*, *Natrix*, *Coluber*, *Elaphe*, *Coronella*, and *Vipera* and the lizard *Pseudopus*) (Khozatskii, 1982; Chkhikvadze, 1988; Redkozubov, 1991). The entire reptile assemblage points to highly mosaic landscapes. There are forms associated with dry and mostly rocky steppe biotopes, ranging even to semidesert stations (*Coluber* and *Eryx*), as well as species indicative of humid forest (*Vipera* and *Elaphe*) and riparian biotopes (*Chelydropsis*, *Sakya*, *Melanochelys*, and *Emydoidea*). According to Redkozubov (1991), the abundance of aquatic turtles, the presence of forms that have a high carapace (like *Testudo černovi*) and that lack fossorial adaptations, sand boas of the genus *Eryx*, and lizards of the *Pseudopus* genus indicate that temperatures were positive all year (p. 15). It is remarkable that the later faunas show an abrupt decrease in the number and diversity of turtles (only marsh forms were retained). The snake fauna lost sand boa *Eryx*, which was a possible result of climatic cooling about 3.5 Ma ago.

A comparison of our data with materials from more western areas shows that the second half of the Cimmerian (second half of the Dacian and beginning of the Romanian, according to the Romanian scale) was marked by coal formation in the Dacian Basin (Andreescu *et al.*, 1986). This points to a warm and humid climate and a considerable role of forests and ulinginous vegetation. Evidently, the climatic aridization in the terminal Cimmerian had a greater effect on the eastern areas.

In sum, it is worth nothing that the area of the northwestern coast of the Black Sea during the time interval 7–3.5 Ma (Pontian–Cimmerian) was a fluviatile, poorly dissected plain with mosaically distributed open and wooded, highly irrigated, biotopes. The climate was warm and humid with positive winter temperatures. Relative aridization is recorded in the early Pontian and terminal Cimmerian.

At the genus level, the systematic composition of small mammalian fauna was generally stable from the early Pontian to the end of the middle Cimmerian. Evolutionary changes in some lineages occurred mainly at the species and subspecies level. At the middle–lower Cimmerian boundary, the fauna lost a number of genera, and during the second half of the late Cimmerian, a number of new genera appeared. The quantitative ratios of dominant groups endured considerable changes.

Many typical Miocene forms apparently became extinct in the fauna of large mammals at some point in the beginning of the Cimmerian. In the mid-Cimmerian, the remarkable appearance of inhabitants of extremely humid wooded biotopes, which are unknown both in earlier and later assemblages, is observed.

Reviewer Yu.B. Gladenkov

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Age of Flysch-Olistostrome Complex in the Eastern Olyutor Range (Southern Koryak Highland)

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In many areas of the Bering Sea coast, flysch-olistostrome deposits crop out along the Olyutor Range. They occupy the lowest position in the range structure, and are overthrust by nappes showing a southeastern vergence, which are composed of island arc and oceanic complexes of the Late Cretaceous and, probably, Paleogene ages. In the north, these deposits are in contact with the Upper Cretaceous and Paleocene flysch deposits of the Ukelayat trough, and they are cut by the Olyutor Peninsula structures in the south.

Presently, most age determinations varying from the Santonian to Danian (*Geologiya* ..., 1987) are measured in the olistoliths in these deposits. Only one dating was obtained for the matrix in the Taman Lagoon area, where it showed the Eocene age (*Geologiya* ..., 1987).

The purpose of the study, the results of which are presented in this report, was to specify the age interval, during which the flysch-olistostrome complex was formed. The flysch-olistostrome outcrops were investigated in the Cape Vitgenshtein–Ayat Lagoon area, the Bering Sea western coast. Figure 1 shows the structural scheme of this area compiled by Chekhovich (1989). Two types of the flysch-olistostrome sequences are defined in this area. The characteristic feature of the first sequences is the presence of numerous olistoliths within flysch members all composed of cherts and oceanic or island arc basalts. Oceanic basalt flows are also present in an original occurrence within the flysch member having the hot contact with the latter. The sequences of the second type are composed of flysch members lacking any olistoliths. The rocks in the sequences of both types are intensely dislocated and show isoclinal folds and cleavage. The cleavage and fold axial planes dip steadily to the northwest and north-northwest. The considered deposits are overthrust by gently sloping nappes of southeastern vergence, which are composed of cherts and volcanogenic rocks of island arc origin. The relationship between flysch sequences of the first and second types is unclear. Cherty olistoliths from the sequences of the first type yield numerous radiolarians, whose age, according to Vishnevskaya, ranges from the Santonian to Maestrichtian (*Geologya* ..., 1987). A number of radiolarian datings were also obtained for the cherty rocks overthrusting the flysch deposits. The ages of these rocks vary from the Santonian to Danian (*Geologiya* ..., 1987).

New age determinations for the flysch matrix in four blocks of the flysch-olistostrome complex are based on nannoplankton obtained (Figs. 1 and 2).

The sequence of block I is of the first type, and consists of rhythmically alternating black and dark gray siltstones and claystones and gray fine-grained volcanogenic sandstones. In the upper part of the section, there are two olistoliths up to seven meters across showing distinct sedimentary contacts with host rocks. One of them is composed of bedded black and dark green cherts, and the other one consists of gray volcanomictic sandstone. The thickness of the sequence is up to 60–70 m. Nannoplankton assemblage is represented by the following species: *Micula staurophora, Watznaueria barnesae, Ceratolithus* cf. *camptneri, Glaucolithus compactus*, and *Cribrosphaera ehrenbergi*, which indicates the Maestrichtian age of the host rocks.

The sequences of blocks II, III, and IV belong to those of the second type. Block II is composed of rhythmically interlayering black and gray siltstones and claystones and gray fine-grained volcanomictic sandstones. The rocks of this block contain a nannoplankton assemblage including *Watznaueria barnesae, Micula staurophora, Prediscosphaera cretacea*, and *Rucinolithus* sp. of Campanian–Maestrichtian age. The thickness of this member is about 20 m.

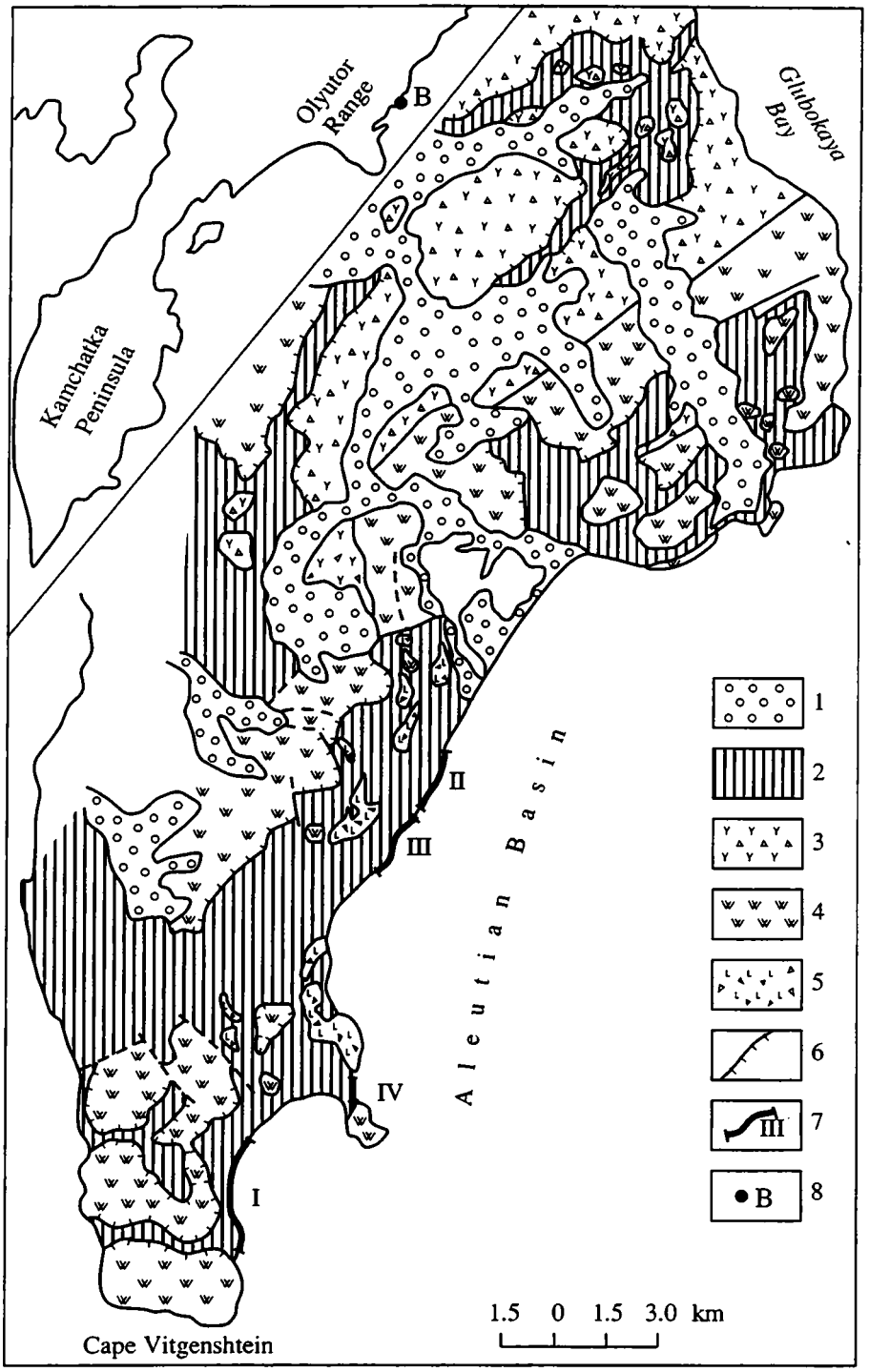

Fig. 1. Location of sampled blocks in the Cape Vitgenshtein structure.
1—loose Quaternary deposits; 2—flysch-olistostrome complex; 3, 4—Senonian deposits: 3—calc-alkaline volcanics, 4—cherty rocks; 5—olistoliths and oistiplaques of oceanic basalt and hyaloclastite; 6—thrusts; 7—sampled blocks of flysch-olistostrome complex and their numbers; 8—location of studied area.

At the base of the sequence of block III, there is a member of alternating gray fine-grained volcanomictic sandstones with a rather distinct gradational structure, and a black siltstone layer. The thickness of sandstone beds is up to 2 m, and that of siltstones is 0.3 m or less. The thickness of the member is about 30 m. Upsection, rhythmically alternating layers (0.2 m thick) of gray sandstone and black siltstone and claystone are observed conformably overlying the previous member. The total thickness of these rocks is 70–80 m.

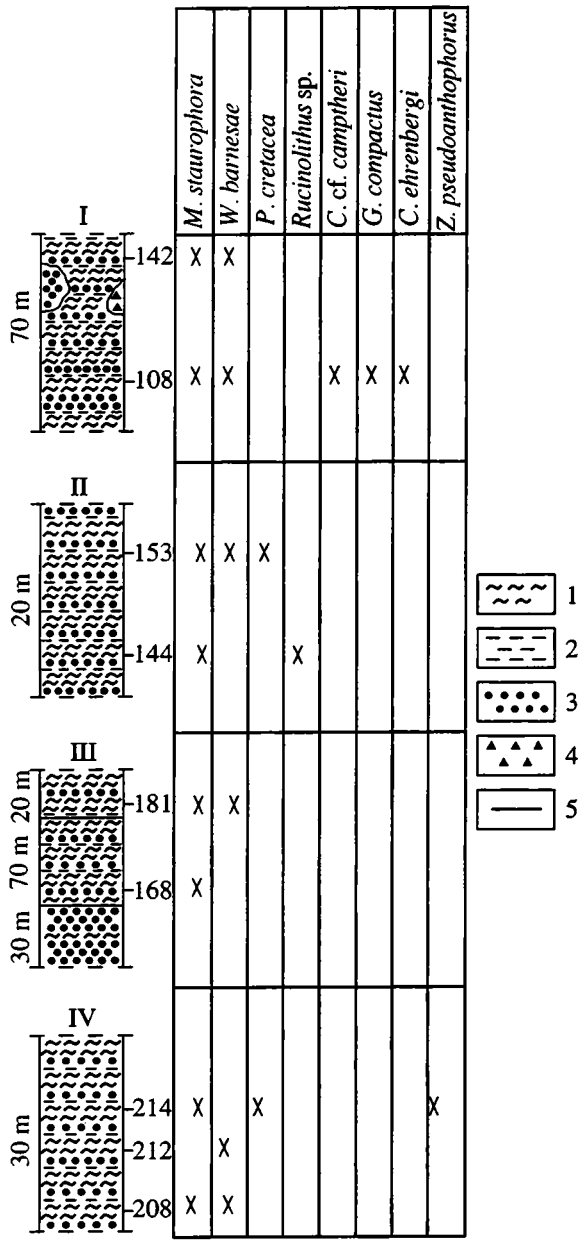

Fig. 2. Lithology of the sequences in studied tectonic blocks of the flysch-olistostrome complex and nannoplankton distribution.

1—claystone; 2—siltstone, 3—sandstone, 4—chert, 5—boundary between members. Numbers to the column right indicate samples with nannofossils.

Black siltstones and claystones dominate higher in the section. The total thickness of the sequence in this block is 120–140 m. Only *Micula staurophora* and *Watznaueria barnesae* were found in these deposits, which refer them to the Santonian-Maestrichtian.

The sequence of block IV is lithologically similar to that of block II, and represented by interlayering sandstones, siltstones, and claystones, all about 30 m in thickness. The impoverished nannoplankton assemblage consists of *Micula staurophora, Watznaueria barnesae, Prediscosphaera cretacea*, and *Zeugrhabdotus pseudoanthophorus* which indicate the Santonian–Maestrichtian age of the host rocks.

Thus, new datings obtained for the matrix of the flysch-olistostrome complex allow the following conclusions to be deduced: (1) the formation of flysch-olistostrome complex of the Olyutor Range was initiated at least in the Late Cretaceous synchronously with the accumulation of flysch deposits in the Ukelayat trough; and (2) flysch sequences of the first (with olistoliths) and second (without olistoliths) types were formed synchronously under different facial or geodynamic environments to be tectonically juxtaposed in the Cenozoic.

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Reviewer I.A. Basov

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